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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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VOLCANISM AND PETROGENESIS AS ILLUSTRATED IN THE HAWAIIAN ISLANDS

BY REGINALD A. DALY

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ABSTRACT

A general view of the chemical diversity of the Hawaiian lavas is offered. They are regarded as derivatives of a primary (olivine-basalt) magma, whose approximate composition is deduced. The differentiation is largely controlled by density. The density of each type of molten, pore-free lava can be estimated closely from its "norm density," as calculated by the Iddings method, the practical accuracy of which is confirmed by a special test. The actual density of the lava in a volcanic conduit is also a function of the weight percentage of juvenile gas, dissolved and in the bubble phase. This gas plays a double rôle, as flux and as inciter of convection in the lava column. As much as 6 per cent of the primary magma in its original home is assumed to be water in solution.

The genetic importance of this gas is particularly illustrated in the case of Hawaiian trachyte. Its consideration is followed by that of the phonolites which in other islands occur in amounts and structural relations like those characterizing the Hawaiian trachytes. Since the volcanic mechanism working in oceanic areas is probably like that working in continental areas, it becomes necessary to account for the fact that in the latter areas rhyolite so often appears at central vents in volumes and general relations essentially like those exhibited by trachyte and phonolite. The explanation offered goes far toward removing a commonly expressed objection to the theory of crystal-fractionation as the chief cause of chemical diversity in igneous rocks.

The picrite-basalt and chrysophyric basalt are most simply explained by the settling of early-formed olivines in the primary basaltic melt, the material of the crystals and melt being thoroughly mixed by two-phase convection in the lava column. "Andesites" and other "intermediate" lavas are attributed to mixing by the same process; the alkaline and melilite-bearing lavas to reaction of the primary basaltic melt with sedimentary material rich in calcium carbonate.

Comparison is made between the primary magmas under deep ocean and continent respectively; it is found that they do not greatly differ and that a fair interpretation of the facts observed at continental centers of eruption strengthens the set of hypotheses erected to account for both volcanism and petrogenesis in the Hawaiian Islands.

In both oceanic and continental areas the chambers in which differentiation, differential refusion, and other reactions with crust-rock take place are thought to be abyssoliths, that is, more or less vertical injections of the basaltic, substratum melt into the earth's crust.

INTRODUCTION

In some respects the Hawaiian archipelago is unique as a region where theories of chemical and mineralogical diversity among lavas can be tested. There much careful mapping of the rock formations has already been done. There, in unrivaled abundance, superior rock-analyses have been supplied by master chemists: Steiger, Ferguson, Bailey, Hillebrand, Keyes, and Washington. Able petrographers have supplied the mineralogical facts needed for the ultimate interpretation of those analyses and then for use in constructing a good theory of volcanism and petrogenesis in general. As in no other field, attack on these problems is favored by the nature of two active craters in Hawaii Island itself—craters which can be approached safely and studied with instruments of precision. There T. A. Jaggar and his associates, A. L. Day and E. S. Shepherd of the Geophysical Laboratory at Washington, F. A. Perret, and others have accumulated a matchless set of relevant data. In the islands field observation is restricted to the tops of the mighty piles of lava, and conclusions regarding the age, conditions, and composition of nine tenths of each pile must be matters of inference. However, there is no evidence that the average composition of the hidden part differs greatly from the average composition of the visible part. In any case the effort to explain the variety of the accessible lavas, by processes that have been operating since these volcanoes lifted their heads above sea, seems justified. The following paper, a record of this venture into the unknown, is based on the writer's field observations of 1909 in Hawaii, Maui, Oahu, and other islands, at a time when the Halemaumau crater in the Kilauea sink was fully active, and on prolonged study of the literature concerning the whole archipelago.

TYPES OF LAVA

The range of chemical variation among the Hawaiian lavas is conveniently illustrated by using Washington's specific names, even though his definitions of "andesite" and some other words are not orthodox. In his five papers (Washington, 1923; Washington and Keyes, 1926; 1928) 97 superior analyses are reported. From the whole series averages for the types as named by Washington have been struck, and the results are entered in Tables 1 to 6. Since it is uncertain as to how much of the water of the analyses is juvenile, the averages have been computed on a water-free basis, and, still further to facilitate more direct comparison of magmatic properties, each has been reduced to total of 100.00 per cent after excluding S, Cl, Cr_2O_3, and other minor constituents. Averages of the corresponding norms (calculated by Washington) are also entered in the tables. The norms for the grand averages (Table 6) were calculated by the present writer.

An outstanding fact is the close similarity of the 23 analyses represented in the first five columns in each of Tables 1 and 2, where olivine basalts, labradorite basalts, and aphyric basalts are included. The grand average of all 23 analyses is stated in column 2, Table 6; it may be referred to as that of olivine basalt. Because the range of composition of the 23 analyses is so small, it seems most unlikely that this

TABLE 1.—*Average analyses: olivine basalts, labradorite basalts, and aphyric basalts of Hawaii Island*

(Water free)

	1	2	3	4	5	6	7
	Pre-Pahala olivine basalt	Aphyric basalt, Kohala	Aphyric basalt, Mauna Kea	Ancient olivine basalt, Kilauea	Ancient labradorite basalt, Kilauea	Mean of cols. 1 to 5	Recent labradorite basalt, Kilauea
No. of analyses in average	2	2	2	2	3	11	4
SiO_2	49.80	48.48	47.41	50.25	51.03	49.58	50.38
Al_2O_3	11.92	13.37	15.10	12.44	13.33	13.19	12.95
Fe_2O_3	4.16	2.48	3.23	1.45	1.25	2.40	1.61
FeO	9.06	9.31	9.22	10.30	9.54	9.49	9.89
MgO	8.60	8.50	7.70	9.64	7.47	8.30	7.95
CaO	10.24	11.02	10.90	10.51	10.76	10.69	11.03
Na_2O	2.24	2.18	2.58	2.21	2.09	2.25	2.35
K_2O	.33	.60	.97	.48	.41	.55	.47
TiO_2	3.45	3.51	2.48	2.35	3.75	3.17	2.92
P_2O_5	.13	.48	.29	.20	.22	.26	.32
MnO	.07	.07	.12	.17	.15	.12	.13
	100.00	100.00	100.00	100.00	100.00	100.00	100.00
NORMS							
Q	4.14	.36		.27	4.50	1.02	.90
Or	1.95	3.33	5.84	2.77	2.58	3.36	2.78
Ab	18.61	18.06	21.09	18.55	17.54	18.94	19.92
An	21.29	24.58	26.58	22.61	25.53	24.30	23.36
Di	22.38	20.91	20.42	23.15	21.15	21.74	24.00
Hy	16.78	19.23	1.62	21.63	18.92	20.42	20.63
Ol	.62	.70	12.94	3.76			
Mt	5.92	3.59	4.64	2.09	1.85	3.50	2.32
Il	6.47	6.60	4.72	4.47	7.11	6.10	5.47
Ap	.34	1.17	.67	.34	.56	.62	.62
Ne			.57				
Norm density	3.186	3.140	3.083	3.132	3.133	3.130	3.132
Measured		2.895 (1 sp.)	3.004	3.043 (1 sp.)	2.908	2.956	2.889

magmatic type is a differentiate from any of the other magmatic types found in the Hawaiian islands. It is much more reasonable to regard the olivine basalt as the material out of which all the other lavas of Hawaii have been made, and for the most part without contamination.[1]

[1] It is of interest to compare the olivine basalt of Hawaii with a fragment from an olivine-basalt flow dredged at the 2000-meter isobath from the top of the mid-Atlantic Swell, during a traverse of the METEOR Expedition (Correns, 1935, p. 289). This specimen, quite fresh and partly glassy, has been analyzed. After reducing to a water-free basis and to total of 100.00, the analysis gave the following percentages: SiO_2, 49.84; Al_2O_3, 16.56; Fe_2O_3, 2.32; FeO, 7.60; MgO, 7.96; CaO, 11.51; Na_2O, 2.64; K_2O, 0.31; TiO_2, 0.78; P_2O_5, 0.08; MnO, 0.19; Cr_2O_3, 0.04; BaO, 0.17. The alumina is about 3 per cent higher than in the average for the Hawaiian basalt, and titanium oxide about 2.3 per cent lower. It is possible that these differences have been exaggerated by a faulty method of analysis. Otherwise this unique specimen from the Atlantic floor is chemically much like the basalt in the Hawaiian islands of the Pacific.

TABLE 2.—*Average analyses: basalts and andesites of Hawaii, Niihau, Necker, Oahu, and Lanai islands*

(Water free)

	1	2	3	4	5	6
	Aphyric andesine basalt, Mauna Loa	Aphyric andesine basalt, Mauna Kea	Aphyric basalt, Hualalai	Olivine basalt, Niihau	Aphyric basalt, Oahu and Lanai	Feldspar phyric basalt, Kohala (2) and Mauna Kea (2)
No. of analyses in average	4	2	1	3	2	4
SiO_2	52.35	50.58	47.90	48.25	51.02	48.30
Al_2O_3	11.63	13.34	15.31	15.37	14.99	15.78
Fe_2O_3	2.10	4.38	5.70	2.47	2.08	3.39
FeO	9.03	8.32	7.65	8.85	8.51	8.28
MgO	7.42	6.31	5.75	8.85	6.78	5.32
CaO	10.14	9.80	10.10	10.74	10.48	10.65
Na_2O	2.26	3.04	3.25	2.57	2.56	3.49
K_2O	.45	.81	.79	.51	.47	.93
TiO_2	4.24	2.68	3.12	2.02	2.72	3.35
P_2O_5	.28	.58	.33	.21	.30	.40
MnO	.10	.16	.10	.16	.09	.11
	100.00	100.00	100.00	100.00	100.00	100.00
NORMS						
Q	8.02	2.54			2.43	
Or	2.50	4.99	4.45	2.77	2.78	5.70
Ab	19.30	25.60	27.27	22.60	21.99	27.89
An	19.56	20.10	24.74	26.92	27.76	24.40
Ne				.94		.77
Di	23.30	19.54	18.65	19.66	18.17	20.53
Hy	15.30	13.65	8.91	4.73	18.49	1.75
Ol			.76	13.51		6.08
Mt	3.08	6.36	8.35	3.55	3.02	4.85
Il	8.00	5.08	5.93	3.81	4.24	6.32
Ap	.67	1.33	.67	.56	.67	1.00
Norm density	3.159	3.117	3.136	3.107	3.058	3.081
Measured density	2.871	3.011	3.001		2.856	2.947

Column 7 of Table 1 shows that in average four recent lavas that emanated from the Halemaumau vent in the Kilauea sink are nearly equivalent with the primary olivine basalt; the reason for this will be suggested later.

Column 6 of Table 2 and all the columns of Table 3 give average analysis and single analysis for types which contrast with one another and also with the corresponding data in Tables 1, 4, and 5. Evidently no simple picture of the derivation of these types from the primary olivine basalt can be drawn.

Table 4 states analyses of four of the most salic lavas. So far as known, these constitute in total no more than 0.1 per cent of the volcanic piles.

TABLE 3.—*Average analyses: andesites rich in alkalies and basalts rich in magnesia*

(Water free)

	1	2	3	4	5	6	7
	Andesine andesites, Mauna Kea	Olivine oligoclase andesite, Kohala	Aphyric oligoclase andesite, Kohala and Haleakala	Andesine andesites, Haleakala	Chrysophyric oligoclase basalt, Mauna Loa	Olivine basalt, Hualalai	Olivine basalt, La Pérouse rock
No. of analyses in average	4	3	3	2	1	2	2
SiO_2	50.49	47.28	52.00	49.04	49.16	46.46	45.99
Al_2O_3	17.53	16.04	16.74	17.96	8.61	13.37	11.57
Fe_2O_3	4.49	4.10	3.07	4.48	2.76	1.88	2.47
FeO	6.35	8.53	7.28	6.59	8.78	9.62	7.19
MgO	4.14	5.81	3.05	2.65	11.96	11.81	16.95
CaO	6.90	7.91	6.42	7.29	10.71	10.47	9.54
Na_2O	4.49	4.33	5.65	5.87	3.20	3.02	2.13
K_2O	1.97	1.48	2.18	2.11	.55	.62	.47
TiO_2	2.69	2.65	2.57	2.69	3.83	2.17	3.08
P_2O_5	.55	1.78	.88	.99	.33	.52	.48
MnO	.40	.09	.16	.33	.11	.06	.13
	100.00	100.00	100.00	100.00	100.00	100.00	100.00
NORMS							
Q	.42						
Or	11.52	8.88	12.42	12.23	3.29	2.48	2.48
Ab	35.53	34.12	41.92	33.31	20.62	17.41	17.41
An	21.70	19.94	13.99	16.60	7.09	20.26	20.26
Ne	1.14	1.14	3.12	8.58	3.07		
Di	6.90	6.55	9.71	10.63	40.48	17.94	17.94
Hy	3.54	.27				4.49	4.49
Ol	5.96	13.78	6.72	4.32	12.46	24.44	24.44
Mt	4.92	5.86	4.40	6.47	3.88	3.44	3.44
Hm	1.12						
Il	5.08	5.01	4.90	5.08	7.02	5.73	5.73
Ap	1.34	4.14	2.30	2.36	.66	2.15	2.15
Norm density	3.000	3.041	2.948	3.000	3.187	3.113	3.147
Measured density	2.794 (3 sp.)	2.878	2.724	2.836 (1 sp.)		3.062	

The most femic of the lavas, namely the picrite-basalts and chrysophyric basalts, have been averaged to give the values of Table 5. Their development, like that of the most salic types, is another subject for theoretical discussion.

DENSITIES OF THE LAVAS

That discussion may well begin with the problem of the densities to be ascribed to the lavas when completely liquid and saturated with water and other volatile matter.

TABLE 4.—*Salic types—single analyses*

(Water free)

	1	2	3	4
	Oligoclasite, West Maui	Anorthoclase trachyte, West Maui	Trachyte, Puu Anahulu, Hualalai	Trachyte, (obsidian), Puu Waawaa, Hualalai
SiO_2	59.86	63.29	62.71	62.63
Al_2O_3	18.89	17.79	18.90	17.56
Fe_2O_3	1.93	5.45	4.35	1.66
FeO	3.76	.08	.10	2.65
MgO	.90	.17	.40	.40
CaO	3.01	1.07	.87	.87
Na_2O	7.35	7.65	6.99	8.34
K_2O	2.89	3.56	4.98	5.06
TiO_2	1.02	.67	.31	.37
P_2O_5	.26	.05	.24	.14
MnO	.13	.22	.15	.32
	100.00	100.00	100.00	100.00
	NORMS			
Q		3.01	1.80	
Or	17.36	20.63	28.99	29.49
Ab	58.58	63.06	58.29	51.90
An	10.08	3.62	2.22	
Ne	2.01			5.40
Ac				4.62
Ns				.73
Di	2.59	.43		2.88
Hy		.20	1.00	
Ol	3.53			3.35
Mt	2.80			
Hm		5.31	4.31	
Il	2.00	.61	.15	.76
Ap	.67		.67	.34
C			1.22	
Ru			.24	
Tn		.78		
Norm density	2.781	2.747	2.724	2.675
Measured density	2.680	2.577	2.763	

A first step is to estimate the density of each type of lava when completely crystallized and devoid of gas-pores. This can be done by the method developed by Iddings (1920). He showed that for an igneous rock which was thoroughly crystallized under heavy cover—a plutonic rock—the density calculated from the norm is greater than the measured density by an amount determined by the porosity reasonably assignable to that rock. The porosity is that expected because of (differential) decreases of volume with the crystallization into a multi-component rock, and because of the development of, and enlargement of, cleavage spaces when such an originally deep-seated rock is exposed at the surface, a place of low pressure. In order to compare norm density and measured density it is necessary to allow for

the content of water. When this is done, the porosities calculated for plutonic rocks are close to those generally accepted for the different species (Daly, 1935, p. 657).

TABLE 5.—*Average analyses: picrite-basalts*

(Water free)

	1	2	3	4	5	6	7	8
	Mauna Loa	Mauna Kea	Kilauea	Halea-kala	Kauai Island	Nihoa Island	Necker Island	Average of 1-7
No. of analyses in average	2	1	3	1	1	1	2	11
SiO_2	48.75	46.44	46.82	43.66	40.67	46.17	45.18	45.93
Al_2O_3	9.91	7.79	8.72	10.35	8.84	10.33	8.56	9.13
Fe_2O_3	1.73	2.40	2.04	3.05	6.00	3.80	4.38	3.05
FeO	9.85	8.89	10.56	10.43	8.26	9.23	8.93	9.64
MgO	17.58	19.68	20.93	14.84	16.38	17.56	17.76	18.36
CaO	7.74	10.63	7.19	12.72	11.81	7.41	8.20	8.73
Na_2O	1.69	1.70	1.42	1.41	3.14	1.93	1.41	1.70
K_2O	.38	.33	.32	.53	1.19	.60	1.15	.61
TiO_2	2.03	1.67	1.72	2.54	2.74	2.57	3.83	2.40
P_2O_5	.22	.34	.14	.30	.82	.25	.37	.30
MnO	.12	.13	.14	.17	.15	.15	.23	.15
	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
NORMS								
Or	2.20	1.66	1.67	3.23	6.38	3.23	6.70	3.33
Ab	14.89	14.02	11.86	5.21	5.66	15.74	12.12	14.57
An	18.19	12.67	16.66	19.64	14.18	17.21	13.70	15.51
Ne				3.67				
Di	14.30	29.98	14.47	32.76	36.76	13.64	19.49	20.47
Hy	26.11	.40	18.96			20.04	10.33	8.34
Ol	17.04	33.81	29.63	23.70	20.71	17.89	23.09	27.64
Mt	2.52	3.44	2.94	4.39	8.80	5.39	6.40	4.40
Il	3.76	3.16	3.24	4.83	5.16	5.15	6.64	4.55
Ap	.66	.66	.34	.67	2.02	.66	.67	.62
Norm density	3.168	3.203	3.204	3.201	3.341	3.202	3.281	3.212
Measured density	3.080	3.164	3.033					3.071 (6 sp.)

For the computation of the norms of the Hawaiian rocks the standard minerals were taken to have the following densities:

Quartz	2.65	Hypersthene	3.30
Orthoclase	2.54	Olivine	3.30
Leucite	2.48	Akermanite	2.95
Albite	2.61	Magnetite	5.00
Anorthite	2.765	Ilmenite	4.73
Nephelite	2.62	Hematite	5.22
Corundum	4.00	Apatite	3.20
Diopside	3.275	Calcite	2.72

In general a solidified lava retains gas-bubbles; in many cases it contains interstitial glass, whose density is lower than that of the rock as a whole; and in all cases it solidifies without the compacting influence of a heavy cover. It is thus clear

TABLE 6.—*Averaged analyses compared*

	1	2	3
	Average of 11 analyses of Table 1 (cols. 1–5)	Average of 23 analyses, including 11 of Table 1 and 12 of cols. 1–5 of Table 2	Average of 4 analyses of recent lavas from vent at Halemaumau, Kilauea
SiO_2	49.58	50.04	50.38
Al_2O_3	13.19	13.47	12.95
Fe_2O_3	2.40	2.65	1.61
FeO	9.49	9.05	9.89
MgO	8.30	7.79	7.95
CaO	10.69	10.49	11.03
Na_2O	2.25	2.43	2.35
K_2O	.55	.55	.47
TiO_2	3.17	3.12	2.92
P_2O_5	.26	.29	.32
MnO	.12	.12	.13
	100.00	100.00	100.00
NORMS			
Q	1.02	1.56	.90
Or	3.36	3.34	2.78
Ab	18.94	20.40	19.92
An	24.30	24.19	23.36
Di	21.74	21.35	24.00
Hy	20.42	18.90	20.63
Mt	3.50	3.71	2.32
Il	6.10	5.93	5.47
Ap	.62	.62	.62
Norm density	3.130	3.127	3.132

why the norm and measured densities of the Hawaiian lavas differ so widely—as illustrated in Tables 1 to 5.

A check on the Iddings method of computing the density which a lava would have had if it were crystallized completely and without voids of any kind is found in a comparison between the norm density at 20° C. (3.090) of the Kilauean "Pele's hair" and its measured density. This glass is practically free from glass pores and crystals. Its measured density at 20° is 2.850. The difference between 3.090 and 2.850 is 9.67 per cent of 3.090, as indicated in Table 7. Table 8 gives for diabase and various pyroxenes and lime-soda feldspars the percentage change of density due to change of state. (*See* Day, Sosman, and Hostetter, 1914, p. 1–39; Dane, 1941, p. 809–818; Daly, 1933, p. 50.) When allowance is made for the porosity of the diabase and for its smaller proportion of the pyroxene molecules and of tita-

ium oxide, it seems right to assume that the Kilauean basalt should show the decrease o approach 9.67 per cent rather closely.

A second check, with result similar in principle, is seen in the case of a basaltic lass derived from a dike near the Kilauean vent (Tables 7, 8).

TABLE 7.—*Basaltic glasses from Kilauea*

	Pele's hair		Dike apophysis	
		Water free		Water free
SiO_2	49.23	49.64	50.66	50.83
Al_2O_3	12.94	13.05	13.64	13.68
Fe_2O_3	1.76	1.77	1.69	1.70
FeO	9.28	9.36	9.33	9.36
MgO	7.42	7.48	7.03	7.06
CaO	11.28	11.38	10.78	10.82
Na_2O	2.64	2.66	2.16	2.17
K_2O	.59	.59	.35	.35
H_2O^+	.74		.27	
H_2O^-	.08		.07	
TiO_2	2.75	2.77	3.69	3.70
P_2O_5	1.21	1.22	.20	.20
MnO	.08	.08	.13	.13
	100.00	100.00	100.00	100.00
NORMS				
Q	.18		4.63	
Or	3.34		2.23	
Ab	22.57		18.40	
An	21.45		26.22	
Di	22.87		21.48	
Hy	18.42		16.79	
Mt	2.55		2.56	
Il	5.32		7.01	
Ap	2.48		.34	
(H_2O)	.82		.34	
	100.00		100.00	
Norm density	3.155		3.090	
Measured density	2.850 (90.33% of 3.155)		2.819 (91.2% of 3.090)	

Having systematically applied the Iddings method and so calculated the norm densities indicated in Tables 1 to 6, the next step is to estimate the effect of raising the temperature of each glass from 20° to that of the maximum temperature of the upper part of each feeding abyssolith. Let this be taken at 1250°. Experimentally found values for the thermal expansion of diabase and diopside are stated in Table 8. A conservative figure for the Hawaiian lavas, namely 40×10^{-6}, will be used. On that basis it appears that at atmospheric pressure the density of any basaltic or

approximately basaltic lava, when molten and free from gas-bubbles, is about 13 per cent lower than that of the same rock when completely crystallized, without voids, and at room temperature. The percentage change would be greater in the case of the more salic lava.

TABLE 8.—*Densities and thermal expansions*

Material	Authority	Density, crystalline at 20°C., measured	Density of glass at 20°C.	Decrease of density, cryst. to glass, both at 20°C.	Density of melt at 1250°	Decrease of density, cryst. at 20° to melt at 1250°	Thermal expansion per 1°, mean at melting point $\times 10^6$ (approx.)
				%		%	
Diabase	Geophysical Laboratory	2.975	2.763	7.15	2.59	12.9	
Diabase	E. B. Dane	2.969	2.78	6.5	2.64	11.1	38
Diopside (synthetic)	E. B. Dane	3.265	2.846	12.9	2.695	17.4	64
Diopside (natural)	E. S. Larsen	3.275	2.830	13.6			
Enstatite	E. S. Larsen	3.175	2.743	13.6			
Augite	E. Deville	3.266	2.835	13.2			
Albite	Geophysical Laboratory	2.605	2.382	8.55			
Ab_1An_1	Geophysical Laboratory	2.679	2.533	5.45			
Ab_1An_2	Geophysical Laboratory	2.710	2.591	4.4			
Ab_2An_7	E. B. Dane	2.71	2.61	3.7			
An	Geophysical Laboratory	2.765	2.700	2.35			
Pele's hair, Kilauea	R. A. Daly and H. S. Washington	3.155 (norm)	2.850	9.67	*ca.* 2.72	*ca.* 13.8	*ca.* 40
Basaltic glass of dike, Kilauea	R. A. Daly and H. S. Washington	3.090 (norm)	2.819	8.77	*ca.* 2.69	*ca.* 12.9	*ca.* 40

Since the liquid lavas within abyssolithic chambers carry water and other volatiles, either free or in solution, the density of each must be smaller than its own norm density. This fact will need emphasis when we come to deal with the subject of differentiation in the lava columns.

SOURCE OF THE VOLCANIC MATERIAL

Petrologists of the present day are nearly or quite unanimous in explaining the diversity of igneous rocks as fundamentally due to crystal-fractionation. They are far from unanimity in deciding on the original liquid from which all the known types have come, though there is some ground for supposing the ancestral magma to have had the composition of the stony meteorites (Daly, 1943). Less speculative is the ascription of many varieties of lava to the fractionation of basaltic liquid.

The source of the once-molten basalt of the lava plateaus, dike swarms, and sill complexes of the continents and of volcanoes dotting the ocean is a question as old as the science of geology itself, still awaits a definitive answer, and probably will long remain in the field of competing hypotheses. A good choice among these is necessarily preceded by correlating a world of geological facts with the facts of the rapidly developing geophysics and geochemistry. For many years the writer has been attempting that correlation and continues to favor the conception that the earth has a true crust resting on a thick vitreous layer, briefly described as the substratum. Twenty-five years ago, before seismology and the study of the mechanical properties of rocks and glasses had been well advanced, it was thought that now the upper part of the substratum is basaltic, world-circling, and of considerable thickness. The writer still finds no reason to abandon that hypothesis as a reasonable basis for understanding igneous action during the long stretch of time between an epoch in the later Pre-Cambrian and the Cretaceous or perhaps still later period—an interval through which the earth is supposed to have been cooling. On the other hand, the results of geophysical investigation seem to show that the basaltic part of the substratum cannot now have a thickness exceeding a few kilometers, or else that it is a discontinuous as well as thin couche situated between the crust and a world-circling layer of vitreous peridotite (wehrlitic?). The reasons for preferring the substratum hypothesis and for changing its original form are given elsewhere (Daly, 1933, p. 42, 189–213, 551; 1938, p. 42–73, 113–142, 174–200; 1940, p. 398 ff.; 1942, p. 57–65; 1943, p. 426–433, 437–452); for lack of space they will not be repeated here.

However, the conditions for the noncrystallinity of the substratum basalt deserve some additional comments. Recently developed theories of the earth's internal energy have one common feature—the assumption that the vertical heat-gradient flattens with increase of depth below the surface. Jeffreys, L. H. Adams, and a few other geophysicists consider the flattening to be so rapid that except for local "pockets" the whole silicate mantle on the iron core of our planet is crystallized and truly solid. That conclusion is based on presuppositions regarding the distribution of radioactivity in the earth and on the mode of solidification and is thought to furnish the best explanation of the proved transmissibility of transverse or torsional seismic waves at all depths in the silicate mantle. These investigators have laid little stress on the rôle of originally absorbed gases in establishing liquidity for the imagined "pockets." The smaller the content of volatile matter, the deeper must be the top of each "pocket" that may be supposed to yield the decidedly fluent lavas of basaltic plateaus and injections. If the melt from a "dry" pocket reaches the surface at fairly great speed, that melt on arrival should have a temperature considerably higher than 1250° C., the liquidus temperature for "dry" basalt. There is no evidence for the eruption of "dry" basalt at such temperature. On the other hand, if the erupted basalt arrives at the surface while still carrying 6 per cent of water in solution, this lava could have the temperature of 1250° and yet be superheated several hundreds of degrees. Here we have a condition which can ease our understanding of the great length and prolonged liquidity of basaltic flows in spite of the rapid loss of radiated heat *en route*.

That the proportion of water and other volatiles in the silicate mantle of the earth is much greater than that characterizing rock specimens in general seems to be a logical deduction from the prevailing theory of planetary origin—namely, derivation from a star-born nebula. According to cosmogonists of our time, that primitive nebula or gas-ball was made chiefly of hydrogen, though of course containing much oxygen as well as all other known elements. Further, those authorities give reasons for believing the primitive gas-ball to have had many times the mass of the existing earth (Jeffreys, 1929, p. 35; Whipple, 1941, p. 240–244).

If those deductions are sound, it is logical to assume the solution of both hydrogen and water in the silicate mantle. Goranson's (1931, p. 481; 1932, p. 234; 1936, p. 257; 1938, p. 78, 90) experiments on granite and albite glasses suggest the possibility of as much as 10 per cent of water in the liquid silicate mantle at the beginning of its long history. Much of the water would be lost during convection in the mantle, especially near its surface, but, as the greater, deeper part was crystallized by pressure, water would there be ejected and in the free state rise and tend to saturate the top of the still-molten part of the mantle.

After the unsinkable sial, essentially granitic, was differentiated at the surface of the earth, convection in the rest of the mantle and thickening of the pressure-crystallized shell in depth were greatly retarded, so that the enrichment of the more superficial shells in water continued for many millions of years. Indeed it is quite conceivable that after a stable crust was formed the still-liquid part of the mantle became supersaturated with water, which therefore in the form of minute bubbles slowly floated up to the base of the crust (Daly, 1943, p. 422, 427).

The risen water-gas, whether freed by pressure-crystallization at great depth or freed during the delayed convection, would act as a powerful flux. If, as might be expected, the convection was periodic, the lower, basaltic part of the crystallized crust may have been periodically remelted. At such times the volume of the mantle would be increased temporarily, and the crust should have been stretched and fissured. Have we in this speculation a clue to the reason why voluminous plateau basalts have been extruded at the widely separated intervals indicated by field geology (for example, the floods of the Jura-Trias, early Cenozoic)? Does not the best available scheme of cosmogony encourage belief that the water-gas freed at depth keeps broad segments, if not all, of the earth's basaltic layer in the molten, albeit highly viscous, state?

ABYSSOLITHS OF HAWAIIAN CHAIN

We return to the specific mystery of Hawaiian volcanism. Here as elsewhere the writer finds the source of heat and of most of the lava in the basaltic part of the substratum, the crust being estimated at about 50 miles or 80 kilometers in thickness, as measured from the general floor of the central Pacific Ocean. With that fundamental premise accepted as a starting point, the speculative stages in the volcanic evolution may be phrased as follows. The primary "basalt" rises through the crust along abyssal fissures, where each temporarily liquid "wedge" forms an "abyssolith" or "abyssolithic chamber." Along the fissure the substratum liquid rises at intervals and builds on the sea floor a cone or (exogenous) dome of solidified

lava, until the column of enclosed liquid comes into hydrostatic equilibrium with the crust plus ocean water of the central Pacific. It looks as if this stage should be reached when the top of the liquid abyssolith is 13,000 or 14,000 feet above sea level (Daly, 1938, p. 119; 1942, p. 60).

The mechanical conditions thus speculatively described accord with the fact that great volumes of relatively homogeneous olivine basalt, primary and little or not at all affected by differentiation, have been poured out at vents in each of the island masses. At other times the primary melt long remained as such in the abyssolithic chamber, there to be differentiated and thus able to deliver surface flows chemically contrasted with the parent olivine basalt.

ENERGY AVAILABLE FOR PROLONGED VOLCANISM

It can hardly be doubted that the upbuilding of Hawaii, Oahu, Maui, or any other of the piles of lava took hundreds of thousands of years. In the case of Mauna Loa the average interval of dormancy greatly exceeds the average interval of activity. For years at a time basalt not far below the surface has kept the temperature of liquidity, in spite of cooling by adiabatic expansion with rise from the feeding "chamber" and in spite of chilling against the walls of that "chamber" and of the conduit above it. How under these circumstances liquidity can be so long preserved is a grave question affecting all theories about the volcanism of the islands, and the abyssolithic hypothesis itself must face it.

The difficulty about heat-supply is particularly evident at Halemaumau, the Kilauea vent, active for years on end. There the highest measured temperature in the boiling lake of lava is 1185°, some 60° below the liquidus temperature of "dry" olivine basalt. The loss of heat by radiation at the surface of the lake is enormously rapid, and yet the high temperature is long preserved at that surface. It seems impossible to explain the fact by assuming the original temperature of the lava to have much exceeded that of the hottest flow from the flanks of Mauna Loa—probably little higher than 1300°. On the other hand, there are good grounds for ascribing the retention of high temperature in the Kilauea lake (and in the Mokuaweoweo lake of Mauna Loa) to the action of the volatiles which so wonderfully vesiculate the lava of the lake. The writer's observations on Halemaumau in 1909 suggested ways in which the volatiles could supply the heat radiated at the crater.

Clearly one cannot hold that the volcano was kept alive merely by free juvenile gas rising from an abyssolithic injection, for to supply the necessary energy the gas, when approaching the surface, would have had a temperature of many thousands of degrees, even after its own inevitable adiabatic cooling.

Something more could be said for a second idea—that some of the needed heat was supplied by exothermic reactions among the volatiles as the basaltic magma rose to levels of comparatively low pressures. The same hypothesis was enthusiastically developed by Day and Shepherd (1913), who came to regard exothermic reactions of the kind as the essential cause of volcanic action. (*See also* Daly, 1911.) The present writer, however, believes that heat of reaction must be a subordinate factor and finds much more satisfaction in his second suggestion: that long-con-

tinued activity at a volcanic vent is in large part due to what he named "two-phase" (gas-liquid) convection. The reality of the process is manifest to anyone who watches the Kilauea lava-lake in full action. When the magma reaches the surface of the lake, a large fraction of the contained gas-bubbles escapes, whereby the average density of the gas-liquid combination is increased. The partially gas-freed melt must therefore sink, its place being taken by the more vesiculated, expanding magma beneath. The transfer of heat to the radiating surface of the lake is thus incomparably faster than could be attributed to the independent streaming of free gas through the vent.[2]

According to the hypothesis final plugging of the Kilauean vent will occur when the supply of gas in the feeding chamber has been sufficiently decreased by two-phase convection. That this stage will be reached in the near future is implicit in the writer's best guess as to the kind of chamber here represented. For several reasons it was suggested that Halemaumau is the residual vent from an isolated, floored, laccolithic body of gas-rich liquid basalt, thrust into the older lava beds of the Hawaiian pile.

In principle the activity at the Mokuaweoweo vent of Mauna Loa is like that at Kilauea, but on a greater scale. Hence the chamber feeding the vent and flanking outflows of Mauna Loa is assumed to be larger or else to have more direct communication with the substratum. Is it too wild a speculation to think of the Mauna Loa feeder as abyssolithic?

It is not easy to suppose that abyssolith to be a vertical dike thousands of meters in width. Nor are there any signs that the abyssolithic chamber has been enlarged by the stoping of the lower part of the earth's crust. Perhaps it is safer to retain Dana's idea that the feeding chambers of the Hawaiian volcanoes are relatively narrow, dikelike openings in the crust—a conclusion derived from the echelon alinement of the islands. To preserve the liquidity needed for prolonged (intermittent) activity, such a limited abyssolith would have to be the locus of repeated upthrusts of juvenile magma from the substratum. The cause of the upthrusts may be readily conceived to be isostatic.

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[2] When Halemaumau is in full activity its cliffy walls are attacked and dissolved at a number of points, with the resulting formation of as many "grottoes." These roughly hemispherical recesses are the scene of specially intense combustion of the juvenile gas in the air of the grottoes and also loci for specially rapid diving of the lava in two-phase convection. To both processes in combination may be attributed a promising explanation of the thermal gradient in the lake during the spring of 1917, as determined by Jaggar (1917, p. 399). He found the temperature at the lake surface, where not crusted, to be about 1000°; that in the "blowing cones" on the benches bordering the lake to be 350° or more higher (due to combustion of the gases); and that in a "grotto" to be about 1200°. He also proved the temperature 1 or 2 meters below the lake surface to be only 950° or lower. Between the 4-meter level and the bottom of the lake, 13 meters below the surface, the temperature steadily rose to its maximum, about 1170°. The relatively low temperature of the surface layer is readily accounted for by radiative cooling; the high tempetature of the bottom layer seems best explained by two-phase convection.

The heat reflected from the roof of each "grotto" lowered the viscosity of the lava within it and thus permitted the gases to escape in special abundance. This had two effects: developing a great heat of combustion of the gases and also causing a net increase of the density of the lava within the "grotto" furnace. The increase of density caused the rapid diving of the lava in the "grotto," where, in fact, this aspect of the two-phase convection was particularly evident for the observer. The grotto-heated lava sank to the bottom of the lake, taking its high temperature with it. At that depth surface radiation could lower the temperature at only a slow rate. If this reasoning is correct, it shows that Jaggar is right in ascribing some of the heat radiated at Halemaumau to reaction of juvenile gases with the oxygen of the atmosphere.

VOLATILES OF THE PRIMARY BASALT

If two-phase convection is a leading cause for persistent volcanism at a given center, the feeding abyssolith must not only have great volume, or intermittent refreshment from the substratum, but also must be rich in volatile matter, chiefly water. No reliable estimate of the proportion of water in the substratum, and therefore the initial proportion in a basaltic abyssolith, can be made by even the most careful study of accessible basaltic, diabasic, or gabbroic bodies. All these were emplaced at relatively high levels in the crust of the earth. On the way up much of the gas was likely to escape.

Since chilling of the basaltic magma proceeds at variable rates, with a correspondingly variable rate of increase of viscosity, the percentage of water finally retained in the rock is, within its limits, highly variable. The percentage of H_2O+ in fresh diabase dikes and sills ranges from less than 1 per cent to about 2 per cent. This amount is so small that it is not surprising to find no evidence of significant hydrothermal action on walls and roofs. "Pele's hair" of Kilauea, a vesicle-free glass chilled at atmospheric pressure, still retains 0.74 per cent of water. Fresh tachylyte yields 2 per cent or more. A specimen from a fresh basaltic flow gives 20 times more water than is given by a specimen from another flow, though both flows cooled at pressure only a little above the atmospheric. Specially hydrous are some of the fresh basaltic flows of the Deccan; in four of them Washington (1922, p. 772–774) found the percentage for H_2O+ to range from 1.88 to 3.09, the percentage of total water ranging from 2.56 to 3.44.

Those analytical results suggest that, after reaching the earth's surface, basaltic magma may contain as much as 2 or 3 per cent of water and may retain that amount even after solidifying. While the magma was still liquid, some of this water may have been freed in the form of minute bubbles and later redissolved as the magma cooled—in accordance with Goranson's (1931, p. 496) proof that the solubility of water in rock glass increases with a fall in temperature.

How much greater should be the content of water in the basaltic substratum, under a minimum pressure of about 23,000 atmospheres? What is the percentage expected in an abyssolith before any volatile matter can escape from it? The answers to these questions must be largely speculative, and yet bold speculation is necessary if the problem of volcanism is ever to be solved. In fact no theory of volcanism can satisfy until it is brought into relation with a problem at least as elusive—the origin of the earth itself.

Recently the writer (1943) has ventured into cosmogony, a field of so many difficulties and competing hypotheses, and has applied tests of the old idea that the earth was condensed from a much more massive, gaseous, star-derived nebula, composed largely of hydrogen plus the constituents of meteoritic stones and irons. According to that conception our planet was early differentiated into iron core and silicate mantle, the latter being liquid for a considerable time and from top to bottom. Initially the liquid was saturated with hydrogen, either free or combined with oxygen—that is, as water-gas. In such a body one would expect two-phase convection to occur at a rate much higher than that of thermal convection, caused by

radiation and cooling at the surface. Thus a surface layer of the mantle, 100 kilometers or so in thickness, would lose much of its water and hydrogen, which at the surface-temperature of incandescence would leave the earth's gravitational field entirely. After a stable crust was formed, emanating water was not lost to outer space but formed its own envelope, the primitive ocean. Below the crust two-phase convection continued, though more slowly than before. Each overturn meant the descent of somewhat dehydrated and slightly cooled, silicate liquid toward the iron core, and at successive levels this liquid was crystallized by pressure. Simultaneously the silicate liquid immediately under the crust was enriched in water-gas. Whether ultimately the basaltic substratum and peridotitic shell beneath it became saturated or supersaturated with water is a question that must remain open until more is known about the solubility of water in molten, basic rocks at high pressure.

From specially designed experiments Goranson (1931; 1932; 1938) concluded that the liquidus temperature of granite is lowered 300° C. if its glass contains 6 per cent of water by weight, and 450° with 7 per cent of water. Extrapolating from one of his curves, it appears that at 23,000 atmospheres, the minimum pressure on the assumed substratum under Hawaii, granite glass could hold in solution nearly 12 per cent of water. Similar experimental data for basaltic glass are lacking, but we may suppose that at 23,000 atmospheres it can carry in solution at least 5 per cent of water and perhaps as much as 10 per cent. Let us assume the actual content of water in the substratum to be 6 per cent—an amount of the same order as that which Hess (1938, p. 326) thinks to characterize the ultra-femic (peridotitic) melts which in large volumes have invaded the roots of major chains of mountains. Let us assume further that the water lowers the liquidus temperature of basalt as Goranson found in the case of granite—namely, 300°. The pressure of 23,000 atmospheres would raise the liquidus temperature by about 300°. The combined effect of both influences is to give an estimated temperature of about 1300° for the top of the vitreous substratum. Allowing for adiabatic cooling, the substratum magma, if thrust rapidly up to the surface of the earth, would be at about 1200°, a temperature closely approximating the highest temperature of the voluminous flows through the flanking fissures of Mauna Loa. Is not such a degree of correspondence significant in judging the amount of volatile matter available for two-phase convection and also for the fluxing of volcanic vents in Hawaii?

The calculations have been made without reference to any possible dissociation of the water. Yet the analysis of the Kilauean gases gives a percentage of hydrogen, and it seems fair to assume the presence of this specially powerful flux in the upper part of each abyssolith.

Petrologists in general think of the original content of water in basaltic magma as far less than 6 per cent by weight. In Chapter 16 of his book on *The evolution of the igneous rocks* Bowen (1928) has argued strongly for the poverty of granitic magma in water—incidentally something to be expected if the present writer's suggestion (1938, p. 167 ff.) that the granitic batholiths are in largest part the product of "pure melting" of the sial is correct. Bowen (1928, p. 301) mentions 0.5 per cent as giving the order of magnitude for the water-content in basaltic

magma. He came to that conclusion because of an observed fact: Glassy and aphanitic selvages of basaltic dikes and sills are seldom vesicular, as they should be if the water-content exceeded 1 per cent or so. Bowen (personal communication) used the same observation as a fatal objection to the idea of important deep-seated vesiculation in an abyssolith—that is, at levels where the basaltic magma, having risen from its original home, was relieved of pressure to the extent of 10,000 to 15,000 atmospheres. But is the criterion valid?

Visible dikes and sills are located close to the earth's surface, far above the top of the source body. We have no means of knowing how fast the magma moved up, and we cannot know what proportion of its gas-bubbles were segregated by gravity during the rise. Probably, however, many would be squeezed out of the melt as it underwent the intense shearing, if not actual turbulence, demanded in the process of upward injection. Similar squeezing-out of bubbles, with perhaps more thorough effect, might be expected in the case of apophysal dike or sill when this too had to force its way through the resistant crust of the earth. But again let us assume that some bubbles still remain in the selvage material of dike or sill. This material arrives at its contact with a temperature of about 1000°, is rapidly chilled to about 900°, and then more slowly to 100° more or less. From Goranson's experiments on granite glass and feldspar glasses it seems probable that the described drop of about 900° could involve the resolution of water-bubbles with mass totaling 1 per cent of the selvage basalt. If the chilling is quick enough, the selvage is tachylytic; if slower, it is crystalline with the grain of an aphanite. Some nonvesicular tachylytes carry as much as 3 per cent of water. Little is known about the content of primary, juvenile water in the crystallized contact-phases of basaltic dikes and sills. It may well average as much as 3 per cent in such selvage material which contains no cavities of the gas-bubble type.

Is it too extreme to assume that the selvage basalt had lost half its initial water-content by the time it reached the high-level contact where chilling immobilized this material? Considering all factors in the case, is it wise to trust the selvage criterion as to the possibility of vesiculation and concentration of gas at depths 10 to 20 kilometers below the loci of accessible dikes and sills?

That vast quantities of superheated steam do collect under plugs temporarily formed at the tops of columns of basaltic liquid is illustrated by the tremendous sudden emissions of steam at volcanic centers. A good example is that of Vesuvius in 1906; during that great eruption Perret (1924, p. 50) passed through a cloud of the erupted gas, found it respirable, and concluded it to be made chiefly of water vapor. Perret himself thought this water to be resurgent but gave no proof that at least some of it was not juvenile.

Finally, the assumption that primary basalt carries only 1 or 2 per cent of water is hard to reconcile with what appears to be the best theory of the earth's origin, for this theory, as we have seen, demands a much higher initial percentage of water (oxidized hydrogen) throughout the earth's mantle.

To facilitate further discussion of the origin of Hawaiian and other types of lava, it will be assumed that the primary basalt, the basalt of the substratum, has, by weight, 6 per cent of its substance in the form of dissolved water.

SEISMIC-WAVE VELOCITIES AND THE SUBSTRATUM HYPOTHESIS

That assumption makes somewhat more serious the difficulty of reconciling the substratum hypothesis with seismological evidence regarding the compressibility and rigidity of the earth-shells between the depths of 70 and 100 kilometers in the earth. At these depths the velocity of the compressional wave set running by an earthquake ranges between 7.9 and about 8.1 kilometers per second, while the associated torsional wave runs at 4.3 to 4.45 kilometers per second. After a thorough study of available experimental data Birch and Bancroft (1942, p. 489) have concluded that a water-free, basaltic or diabasic melt at the depth of 60 kilometers would transmit the two waves at rates no higher than 6.5 and 3.6 kilometers per second respectively. They make it probable that dry basaltic glass of the oceanite type, with its relatively low percentage of alumina and high percentage of magnesia, would, at the depth of 80 kilometers, give respective velocities no higher than 7.0 and 4.0 kilometers per second. Six per cent of water in solution would lower even these values. It thus appears that our supposed layer of vitreous basalt below the Hawaiian Islands must be thin and that the apparent wave-velocities at the top of the substratum would have to be controlled by the elasticity of the immediately underlying peridotitic material.

Unfortunately no experiments have been made on the elastic properties of such ultra-femic material in the glassy state. We can, however, get some ideas on the subject from the results of experiments on olivine, pyroxenes, and their associates in known peridotitic rocks. For pyroxenes the decrease of density with melting is much greater than for basalt or diabase; probably olivine would show a similar relation. It seems likely, therefore, that the density of the peridotite glass would be only a few per cent greater than that of the basaltic glass at the same pressure and temperature. On the other hand, the cubic compressibilities of crystalline olivine and pyroxene at high pressure are respectively about 30 per cent and 15 per cent lower than that of crystalline basalt. It seems reasonable to think that there would be comparable relations between vitreous peridotite and vitreous basalt. Water in solution to the extent of 6 per cent by weight would lower density and increase compressibility, but probably in each instance by only a small amount. Making what appear to be the best conjectures as to densities and compressibilities, the writer has calculated the velocity of the compressional wave in a vitreous melt composed of 62 per cent of olivine, 31 per cent of pyroxene, and 6 per cent of water, at 1300°, and under the pressure of 25,000 atmospheres. The result is 8.0 kilometers per second. At the pressure corresponding to the depth of 100 kilometers in the earth the velocity would approach 8.2 kilometers per second. Needless to say, this violent and somewhat arbitrary use of meager data is merely suggestive and in no sense demonstrative. The net result of the whole speculation is rather to illustrate a degree of insecurity for the substratum hypothesis until it is further tested by the geophysicists.

If hydrogen is important among the dissolved volatiles, the total weight percentage of these could be smaller than 6 and yet theoretically permit similar estimates of temperature and wave-velocity in the substratum.

The described picture of the outer earth-shells cannot of course be retained if the proved transmissibility of the torsional seismic waves at all depths in the earth's

silicate mantle depends on essentially complete crystallinity for this thick mass of material. However, the torsional wave can be transmitted through a magma if its viscosity is sufficiently raised by pressure. Whether the pressure of 23,000 atmospheres, corresponding to the weight of the crust under the Pacific ocean, could give the high effective rigidity at the depth of 80 kilometers is unknown. Nevertheless, while admitting the danger of fundamental error, the writer continues to favor the substratum hypothesis because, better than any other yet offered concerning the nature of the earth's interior, it eases explanation of isostatic adjustment, orogeny, the development of batholiths and eruptive sequences, and igneous action in general. The connections with these problems has been shown elsewhere (Daly, 1933; 1938; 1940; 1942).

Some petrologists and geophysicists regard the seismological evidence as fatal to the substratum hypothesis and prefer the idea that the visible lavas of the world have risen from separate, floored "pockets" of liquid in the otherwise crystallized and therefore effectively rigid mantle surrounding the earth's core. This competing hypothesis has its own burden of uncertainty; no one has ever suggested a satisfactory reason for such localization of liquidity. Nor is the "pocket" idea a postulate by means of which we can readily account for the above-mentioned, major readjustments of the material in the mantle.

VESICULATION OF AN ABYSSOLITH AND ITS EFFECTS

Whether the volcanist prefers the substratum hypothesis or the "pocket" hypothesis, he cannot avoid the difficulty of explaining the long life of an Etna or Mauna Loa. The required transfer of heat to the surface is not likely to be accomplished by two-phase convection alone. Although at each crater much water gas, hydrogen, and other volatiles escape, with consequent increase of density in the unvesiculated part of the lava, this increase is quite moderate. The resulting density is only slightly greater than that of the unvesiculated magma at the base of the molten column above substratum or "pocket."[2]

Moreover, the convective potential within the lower half of the lava column must also be weak. This conclusion can be read out of Goranson's (1936) curve for the relation between pressure and the solubility of water in albite at 1200°. Table 9 gives values taken from his curve or calculated from his equation for the curve.

It appears that a drop in pressure from 23,000 atmospheres to 10,000 atmospheres lowers the solubility of water in basaltic glass by less than one tenth of the original value. At the deeper levels of an abyssolith any vesicles formed by the change of pressure as the lava rises in an abyssolithic fissure would not only be small but also relatively few in number.

For two reasons, therefore, two-phase convection in the whole column of magma would be slow. Accordingly it seems necessary, as already noted, to think that the upward transfer of heat depends also on sufficiently numerous, upward thrusts of unchilled magma directly from the source of the magma.

On the other hand, the very fact that little gas is released from solution until the

[2] An order of magnitude for the difference of density is suggested by Tilley's (1922) study of rhyolite glasses with different percentages of water in solution.

substratum liquid has risen through half the height of an assumed abyssolithic mass means that convection should be lively in the uppermost 30 or 40 kilometers of the lava column. The convective stirring should be most lively at its top, the place of escape for the rapidly expanding, free gas. Before leaving the silicate melt the vesicles are both abundant and comparatively large. Hence the upper part of the

TABLE 9.—*Weight percentages of water in albite glass at 1200° C. when saturated*

Pressure in bars	Percentage of water in solution
0	0
1000	4.4
2000	6.3
3000	7.2
4000	7.9
5000	8.2
10,000	9.6
15,000	10.0
20,000	10.2
25,000	10.375

Our olivine-basalt glass would probably give a curve which is different and yet would show similar flattening as the pressure increases.

abyssolithic magma has special characteristics. These are important and deserve attention.

First, we note that, if the gas bubbles total only 1 per cent of the weight of this high-level lava, the density of the two-phase fluid would be considerably more than 3 per cent less than the same silicate melt if it were at the same temperature but devoid of vesicles. If the density of the latter be 2.7, vesiculation to the degree described would give a net density no greater than 2.6. More pronounced vesiculation would have still greater effect on the over-all density. It therefore seems clear that the sinking of crystals and also stoping of loose blocks from roof or walls of an abyssolith are hastened by the vesiculation.

Second, it is worth considering the dynamic power of a large mass of the gas collected at the top of an abyssolith on its way up from the depths. Long ago Fisher (1889, p. 279–281) showed how such a "head" on an abyssal injection would greatly aid in the opening of the abyssal fissure itself. On the same principle the injection of offshoots in the form of dikes and sills becomes easier to understand.

The gas released from the primary magma has a third effect. Because of its low density the gas tends to accumulate at the top of the magma column. It there fluxes the walls of the abyssal fissure, which is thereby enlarged locally into a more or less cylindrical pipe. This pipe is henceforth the preferred passage-way for new extrusions of primary magma and also the place for lively convection. The development of just such a pipe is well illustrated in the canyon walls of the Iao valley in West Maui, where a once-active vent was fluxed through a thick series of older lava beds. An open pipe at the top of Hualalai seems to be an analogy. There is some evidence that the vent whence came the many lava fountains of Halemaumau itself was a vertical pipe. (*See* Daly, 1911, p. 98, and *compare* Noble *in* Washington,

1923, p. 121.) Other analogies are probably represented by the feeders of the extinct pit-craters of the Puna (Kilauea) district in Hawaii Island.[4] In none of these cases was the vent opened by explosion of the diatreme type. At many other volcanoes powerful explosion has widened the eruptive throats, but even there the cylindrical form at deep levels may well have been due to gas fluxing.

The fluxing of the conduit walls means assimilation by the active magma, whose composition is more or less changed. If both temperature and the supply of gas have high values for a long period of time, the assimilated material may be an average of the rock that constituted the vent walls to great depth. If this average is the same as that of the somewhat gas-freed primary magma, the active lava in the vent will closely resemble the primary magma. This theoretical deduction seems to account for the small differences between the respective oxide percentages of the primary olivine basalt of Hawaii and the recently active lava of Halemaumau at Kilauea. (*See* columns 2 and 3 of Table 6.)

The described fluxing of the conduit walls has special significance for our main question—the cause or causes for the diversity of chemical types represented by the Hawaiian lavas and inferentially by the lavas in every volcanic region. The solid rocks enclosing the liquid columns of Hawaii are almost entirely igneous and solidified by degrees. From each of those older magmas a crop of high-melting crystals was succeeded by a crop of low-melting crystals. Fluxing of the conduit walls is likewise progressive but in the opposite sense. Therefore the gas-enriched top of the conduit magma tends to become, and on account of density-relations to remain, more salic than the melt deeper in the column. Any high-melting constituents of the disintegrated wall rock, enclosed in the lava column, sink in it and, if redissolved in depth, increase the density of the liquid at the deeper levels. The effect is to increase the vertical heterogeneity of the lava column, gravity separating the contrasted products of refusion as well as the products of incipient crystallization.

But for a long period of time that double tendency toward gravitative differentiation is largely offset by two-phase (gas-liquid) convection. Its vertical currents continue to mix the lighter and heavier melts in the column more or less perfectly, especially in the upper half of the abyssolith. The mixing goes on until the liquid at the top of the column becomes rather decidedly salic and hence of density so low that this part of the lava column can no longer partake of the gas-liquid convection. The critical stage is reached when the upper, more salic liquid attains a silica percentage of about 58 or nearly that of Washington's "oligoclasite." Before that stage is reached convective mixing with the less modified and more femic liquid gives melts of differing composition, but each with less than 58 per cent of silica. In fact, among the rock analyses compiled by Washington none shows a silica percentage between 53 and 58. Barth (1939, p. 65), not allowing for convective mixing, found this gap in the chemical series "hard to understand."

ORIGIN OF THE TRACHYTE

The writer had no opportunity to study the field relations of the Hawaiian trachytes in detail. Observations by other visitors to the islands seem to show

[4] Washington (1917, p. 278; Washington and Keyes, 1926, p. 380) adopted the gas-fluxing hypothesis in explanation of vents on Etna and Stromboli.

that these salic bodies are homologues of the trachytes in Tutuila (Samoa), Ascension, and Saint Helena islands, all of which the writer has mapped since 1909.[5] In all three islands the salic bodies were accumulated at the tops of the respective lava columns. In many instances the rise of the trachytic masses, characteristically in the form of endogenous (crater-filling) domes, through the great piles of olivine-basalt beds was preceded by powerful explosions which yielded breccias, largely or wholly basaltic in composition. Evidently each lava column was headed with a mass of hot gas assembled above the salic lava and presumably held in by a plug which had formed during a period of dormancy.

Plugging and dormancy mean a temporary fall of temperature in the upper part of the lava column. The cooling leads to advanced crystal-fractionation, with the gravitative separation of early crystals from salic liquid, which accumulates just beneath the gaseous "head" of the column. Still deeper, two-phase convection continues, though necessarily slowed down, and tends to distribute the sunken crystals more or less evenly through the liquid at depth. Meantime there is steady thickening of the unsinkable trachytic melt. The relative amount of the salic differentiate at the top of the column is much increased by the selective, gas-controlled refusion of the walls and of the temporary plug. This effect, strongly supplementing that of direct crystal-fractionation, seems capable of accounting for the homogeneous domes of trachyte that rose in so many craters of the three islands. In most cases the cooling of the upper part of the lava column had gone so far that all activity ceased at these particulars centers, though it was continued at other localities in the respective island masses. To the writer it seems likely that similar conditions and sequence of events may be postulated for the bodies of trachyte in Hawaii.

In this connection a determination by Barth (1931, p. 395) is significant. He made a rough analysis of the groundmass feldspar in a holocrystalline basalt from Hiva Oa of the Marquesas archipelago and found about 28 per cent of $Ab_{93}An_{07}$ and 72 per cent of alkali feldspar, the latter consisting of orthoclase, albite, and anorthite in respective percentages of 56, 41, and 3. Schairer and Bowen (1935, p. 328) have shown that such mixed crystals melt at relatively low temperatures. Thus it has been once more proved that potash is largely concentrated in the residual liquid of a basalt as it crystallizes. Chemically the groundmass assemblage of minerals in the Hiva Oa rock closely resembles the anorthoclase trachyte of Hawaii and also the first liquid which may be expected to form during the refusion of the basaltic walls of a volcanic vent in the islands.

In favor of the foregoing explanation of the trachytic bodies is the fact that apparently the volume of none of these exceeds a small fraction—perhaps 1 or 2 per cent—of the volume reasonably to be assigned to its parent abyssolith. This consideration is important whether the trachyte represents liquid residual from crystallizing pure basalt or liquid residual in the crystallization of basalt contami-

[5] *See* Daly, 1924; 1925; 1927. Barth (1931, p. 521–522) lists other Pacific, dominantly basaltic islands containing small bodies of trachyte: Clipperton, Pitcairn, Society, the Marquesas, San Felix, and Juan Fernandez. All these trachytes, like those of Hawaii and Samoa, "belong to the soda-trachyte group (the only possible exception being the 'latite' of the Marquesas Islands)."

nated with crust-rock, itself of basaltic composition. The corresponding enrichment of the lower part of the abyssolith in mafic, sunken crystals is also small. Convectively mixed with the basaltic liquid at depth, they should not form any large mass of ultra-mafic lava capable of eruption.

TABLE 10.—*Chemical comparison of picrite-basalt and olivine-enriched primary olivine basalt*

	1	2	3	4
	Primary olivine basalt	Average olivine crystal	Mixture of 1 and 2, reduced to 100 per cent	Average picrite-basalt of Hawaii
SiO_2	50.04	40.33	47.63	45.93
Al_2O_3	13.47	.42	10.21	9.13
Fe_2O_3	2.65	1.13	2.27	3.05
FeO	9.05	10.39	9.37	9.64
MgO	7.79	47.20	17.65	18.36
CaO	10.49	.16	7.91	8.73
Na_2O	2.43	.12	1.85	1.70
K_2O	.55	.06	.43	.61
TiO_2	3.12	.08	2.34	2.40
P_2O_5	.29		.22	.30
MnO	.12	.11	.12	.15
	100.00	100.00	100.00	100.00

ORIGIN OF THE PICRITE-BASALT

Theoretically the Hawaiian trachyte represents the final, residual liquid resulting from the crystal-fractionation of our primary olivine-basalt magma. In greatest chemical contrast with it is picrite-basalt (including chrysophyric basalt), and it seems simplest to ascribe this mafic lava to the first stage in that fractionation. The microscope shows olivine to have been the first abundant mineral to form in the cooling primary melt. On account of superior density it sinks to levels in the lava column where the temperature has not fallen so much and where vesiculation by freed gas is still important. At those levels the over-all composition becomes that of the magnesia-rich picrite-basalt. In principle Table 10 illustrates the case. Column 1 gives the composition of the primary magma (calculated as water-free); column 2, the average analysis of three olivines from Kilauean and Mauna Loa flows (Aurousseau and Merwin, 1928, p. 560); column 3, the composition expected after adding one part of the olivine by weight to three parts of the primary magma; and column 4, the average analysis of the picrite-basalt of the Hawaiian islands. The similarity of columns 3 and 4 seems to show the adequacy of the process so far as gross composition of the picrite-basalt is concerned.[6]

[6] Quite recently Macdonald (1944, p. 177) has described the remarkable contrast between the lavas erupted in 1840 at two points on a long fissure traversing the region of pit craters east of Kilauea. One of the flows, emanating between the 2650-foot and 3100-foot contours, is basalt comparatively poor in olivine. The other (Nanawale Bay) flow, emanating between the 750-foot and 850-foot contours, is strongly chrysophyric, with percentages of phenocrystic olivine between 29 and 35. Macdonald attributes the abundance of these phenocrysts to crystal-settling. Steiger found the chrysophyric rock to carry 19.96 per cent of magnesia. The higher-level basalt has not been analyzed. At least its magnesia content should be determined, to furnish a promising test of this particular application of the crystal-settling theory.

However, one characteristic of these strongly mafic flows deserves special attention. In each flow the conspicuous olivines are not grouped or "rafted" together but are distributed through the rock with a high degree of uniformity. Two questions arise. Do these crystals directly represent the original olivines that came down from cooling basalt higher up in the lava column? Or are they new phenocrysts, individualized after the sunken olivines had been resorbed at the deeper levels? Such individualization would occur if the deeper-lying lava later lost some temperature or free gas, or both, and would be expected to progress after the mafic lava flowed out, to chill under the air. In either case the even distribution of the visible olivines could readily be attributed to two-phase convection which would not fail to mix the material of the sunken olivines all through the deep-lying primary magma.

ORIGIN OF THE "ANDESITES" AND OTHER INTERMEDIATE TYPES

Two thirds of the 97 analyses stated in Washington's papers, like the averages of Tables 2 and 3, have silica ranging within narrow limits—between 47 and 53 per cent—but show relatively great differences in the proportions of the other oxides. For this, simple crystal fractionation under gravity, considered by itself alone, offers no explanation. Doubtless the same would still be true if we had 1000 or more analyses in hand. The gravitative or other separation of successive crops of crystals from a primary magma is inadequate for the evolution of these many lava types which in a sense are "intermediate" between trachyte and picrite-basalt. On the other hand, a good solution to the problem is found if we assume the products of crystal-fractionation to have been more or less steadily accompanied by two-phase convection and therefore mixing in an abyssolithic body of primary, olivine-basalt, magma, especially in its upper part. That convection of this type was possible is indicated by the small range of norm densities for all the Hawaiian lavas that are more salic than olivine basalt and more femic than oligoclasite. (*See* Tables 2 and 3.)

Here again, study of the activity at Halemaumau, Mokuaweoweo, Matavanu (in Savaii of the Samoan group of islands), Etna, and Vesuvius, at all of which two-phase convection is manifest, shows that the crystal-fractionation in a volcanic column of lava is much more complicated than that in a laboratory crucible.

ORIGIN OF THE ALKALINE LAVAS AND OF THE MELILITE-BEARING LAVAS

Most of the Hawaiian rocks are unsaturated, and among them is a number of types belonging to the alkaline class as defined by Shand. Some details concerning them are to be found in Barth's paper (1931, p. 403, 508–516, 526–529), which gives references to other sources of information. To the tephrite-basanite group he assigns a limburgitic basalt of Niihau island, a nepheline andesite-basalt of Haleakala (East Maui island), and an olivine-oligoclase andesite of Kohala (Hawaii island). To his "pacificites" (the basantoides and tephritoides of Lacroix) he refers a limburgitic basalt, an andesine andesite, and a nepheline-andesine basalt of Haleakala; an andesine andesite and a chrysophyric oligoclase basalt of Mauna Kea; an olivine

TABLE 11.—*Analyses of an alkaline lava and of melilite-bearing lavas*

	1	2	3
	Nepheline basalt, Necker	Nepheline-melilite basalt, East Oahu	Melilite-nepheline basalt, Kauai
SiO_2	47.89	36.34	37.50
Al_2O_3	16.21	10.14	9.12
Fe_2O_3	5.39	6.53	5.59
FeO	4.99	10.66	8.81
MgO	3.19	10.68	13.72
CaO	7.79	13.10	13.85
Na_2O	7.73	4.54	2.69
K_2O	1.38	1.78	.63
H_2O^+	.62	1.00	2.35
H_2O^-	.10	1.00	1.05
TiO_2	4.81	2.87	3.21
P_2O_5	.19	1.02	.90
MnO	.06	.20	.15
CO_2	none	.15	.27
S		.04	Tr
Cl			.05
Rest			.30
	100.35	100.05	100.19
NORMS			
Or	8.34		
Lc		8.28	2.62
Ab	29.34		
Ne	19.31	21.02	12.21
An	5.56	1.67	11.12
Di	17.68	17.39	31.04
Ol		20.23	18.56
Wo	4.18		
Ak		11.95	3.92
Mt	2.09	9.51	8.12
Hm	4.00		
Il	9.12	5.47	6.08
Ap	.34	2.35	2.12
Calcite		.31	.60
Norm density	3.078	3.226	3.250

basalt of Hualalai; and an olivine-oligoclase andesite of Kohala. True phonolites are extremely rare in the Hawaiian archipelago. None is listed by either Cross or Washington. According to Barth the only analyzed representative is indicated by a nepheline basalt composing the chilled margin of a dike in Necker island (Table 11, col. 1).

Table 11 gives the analysis of a nepheline-melilite basalt from East Oahu (column 2) and that of a melilite-nepheline basalt from Kauai island, the latter quoted from Cross (1915, p. 17, 22). For a theoretical reason these and other melilite-bearing

rocks—the latter from West Maui, West Oahu, and Niihau—are to be considered in relation to the genetic problem offered by the phonolitic nepheline basalt of Necker island.

Both Cross (1915, p. 93) and Washington (Washington and Keyes, 1928, p. 218) preferred to regard the alkaline types, like all others except olivine basalt, as differentiates of this primary magma. Cross ended his memoir with the statement that "the processes of this differentiation are still problems for investigation. To me they appear to have acted mainly on the liquid magma. Movement, under gravity, of crystal particles may have played a part." In the case of picrite-basalt Washington favored Bowen's explanation by the gravitative settling of olivine crystals from high-level basalt into deep-level basalt, and in general seems to have been satisfied with the crystal-fractionation of uncontaminated olivine-basalt melt as accounting for all the lava types of the Hawaiian islands. Barth (1931, p. 528) is of the same opinion.

All three writers concluded that reaction of the primary basaltic liquid with carbonate rock could have had little to do with the development of volcanic species in the Hawaiian islands or elsewhere in the ocean. Assuming the only material that could be conceived as entering into the reaction was coral-reef limestone, they came to that belief. There is in fact no evidence that thick masses of coral limestone are intercalated among the lavas of all the enormous piles represented by Hawaii, Tahiti, Saint Helena, and other thalassic islands where feldspathoidal rocks are found. But coral reefs are by no means the only carbonate-rich formations deposited in the open ocean.

Many of the deep-sea islands are geologically young, having begun their upward growth from the bottom probably in post-Miocene time. Some of them are still volcanically active. Many more proclaim relative youth by standing high above wave-base. For long periods before the initiation of eruption the general floor of the ocean had been receiving deposits of calcareous oozes and red clay. Since an early epoch of the Cretaceous, if not earlier, Globigerina entered the general plankton, and it seems altogether probable, therefore, that Globigerina ooze was accumulating for something like 100,000,000 years before those younger volcanic piles began to grow up. That sedimentary layer was buried under each pile of lava. At present Globigerina ooze covers about 40 per cent of the sea floor. Another 40 per cent is now covered with red clay, which in the Pacific area may have been thickening for more than a billion years—from the late Pre-Cambrian onward.

Kuenen (1941, p. 161; 1937, p. 457) has made a courageous attempt to estimate the average total thickness of each of the kinds of sediment, arriving at results which have surprised the geological profession. He used several methods; only one will be described here. Schott (1937, p. 129) found the rate of post-Glacial deposition of Globigerina ooze at different localities in the Atlantic basin to vary from 0.53 centimeter to 2.13 centimeters per 1000 years, the average rate being 1.2 centimeters per 1000 years. Piggot and Urry (1942, p. 1204) estimated the present rate in the Cayman trough of the Caribbean Sea to be 4 centimeters per 100 years or 4 times the average rate they assign to the Pleistocene Glacial period. During this same period they estimate the average rate of deposition for the red clay of the Pacific

0.7 centimeter per 1000 years. On the assumption that the various estimates of rate are of the right order and are not far from average rates in the past, Kuenen found the total average thickness of Globigerina ooze, when supposed to be compacted, pore-free, and water-free, to be 120 meters (the product of deposition since a mid-Cretaceous epoch); to the red clay he assigned a total average thickness of 2500 meters, also arbitrarily supposed to be pore-free and water-free. The results are startling, but who can now say that he failed to give us the right orders of magnitude? If we accept his figure for the average thickness of the ooze and recall the proof of Piggot and Urry that the rain of planktonic shells varies greatly from region to region in the oceanic areas, actual thicknesses of 500 or 600 meters of ooze material, regarded as pore-free and water-free, would be thought reasonable.

It is of course well known that Globigerina ooze is rich in calcium carbonate. The CHALLENGER chemists reported this compound at 53 per cent of six dried samples, by weight. They also found that eight samples of red clay, taken from depths of from 2300 to 2750 fathoms, yielded 31 per cent of the carbonate. In each case the average content is much less; nevertheless Kuenen's discussion tends to support the idea that the volcanic piles of the open ocean do rest on relatively great masses of calcium carbonate—masses much greater than those represented by coral reefs.

And where so little is known about the origin of phonolite and other feldspathoidal rocks in oceanic volcanoes, our drastic speculation may well enter still another field. In a number of publications the present writer has proposed a possible explanation of the general lack of calcareous fossils from marine animals in the Cambrian and Pre-Cambrian sedimentary terranes, such as the thick Belt series. According to the suggested hypothesis, the ocean of those early times was kept nearly free from calcium because of its steady precipitation by the ammonia generated during the bacterial decay of (soft-bodied) animal carcasses fallen from the surface layer of the ocean. Is it out of the question that many hundreds of meters of limestone were in this way deposited on the sea floor long before some post-Cambrian epoch when bottom scavengers, in numbers sufficient to inhibit the ammonia reaction, were evolved? (*See* Daly, 1907, p. 93; 1909, p. 153.)

Obviously it will be long before these guesses about Pre-Cambrian, Paleozoic, and Mesozoic sedimentation on the floor of the open ocean can be adequately tested. Yet perhaps enough has been said to cast doubt on a supposedly fatal argument against the limestone-reaction hypothesis in explanation of oceanic eruptives which are either alkaline or carry melilite (with akermanite, gehlenite, and $CaSi_3O_7$ molecules), lime-rich pyroxenes, and titanite in some abundance.

The top of any of the imagined thalassic deposits was about 4000 meters below sea level to start with and became still deeper as the crust yielded to the weights of the lava piles. The carbonate buried under these piles ultimately came into contact with molten, abyssolithic basalt at levels where much gas was freed from the melt and where two-phase convection was lively. Under such conditions strong reaction between the primary melt and the carbonate-rich sediment would be expected. Of course reaction with any coral-reef or other calcarous material intercalated in each pile of the older lavas would have similar effects.

Space will not here be taken to go into details concerning the reactions; on that

subject Shand's writings (1943) are the most illuminating. However, one item may be mentioned. From the data of Barth (1931) 17 pyroxenes from "lime sub-basalts" of the Pacific give an average of 25 per cent for the diopside molecule, while 10 "pacificites" (carrying occult nepheline) have an average of 42 per cent for the same molecule.

It is worth noting that, according to the hypothesis, any salic differentiate of phonolitic tendency rose from levels relatively high in each abyssolith, so that the space where both reaction and differentiation went on was but a small fraction of the volume of the abyssolith as a whole.

And, finally, the rôle of the resurgent carbon dioxide, added to the juvenile carbon dioxide and other gases, should also be considered when dealing with the causes for the enrichment of the top of the abyssolithic magma with the alkalies.

The phonolitic bodies of the deep-sea islands closely resemble the trachytic bodies in volume, structural relations, and time relations. Since the densities of the corresponding liquids are nearly identical, it is logical to assume the segregation of the phonolitic material to have been accomplished in the way already deduced for the trachytes—namely, by gravitative separation upward.

On the other hand, the melilite-bearing rocks can hardly be referred to simple gravitative differentiation in any magmatic column. Their content of soda and potash rivals or considerably exceeds that of the primary basalt, while they carry much more lime, magnesia, and iron oxide than the basalt. Simple gravity may be expected to cause downward segregation of the abnormal melts if they hold no more volatile matter than that characterizing any of the types represented in Table 11. It is, however, conceivable that before eruption these melts were vesiculated enough (possibly by resurgent CO_2 as well as by juvenile gases) to be able to partake in two-phase convection. The convection currents would tend to mix strongly alkaline materials with material rich in iron, lime, and magnesia. With the volatiles in sufficient abundance the result would be a liquid from which the melilite, olivine, augite, iron oxides, and nepheline of the melilitic lavas finally crystallized. That these lavas were formerly completely liquid (apart from enclosed gas-bubbles) is suggested by their aphanitic or fine-grained character. Partly for this reason the writer doubts that Bowen (1928, p. 259 ff.) has succeeded in his attempt to explain melilitic rocks as derivatives of pure basalt, and finds more to be said for an explanation founded on the limestone-syntectic hypothesis.

The second hypothesis was worked out in the effort to account for the common development of lime-rich crystals, including melilite (akermanite-gehlenite) in alkaline rocks and in closely associated rocks of igneous origin. The general idea would be more widely acceptable if it were supported by the results of properly designed experiments in the laboratory. These should be made on basaltic, granitic, and other subalkaline melts well supplied with water, carbon dioxide, and other volatiles, to represent the juvenile as well as resurgent volatiles of Nature. The experiments would be all the more compelling if the artificial products of reaction and differentiation could be subjected to mixing during experimental runs.

COMPARISON WITH CONTINENTAL VOLCANISM

The abyssolith (crust-substratum) hypothesis has been applied to the volcanic problems of the mid-Pacific. If there valid, it should be applicable to the same problems in the broad continents and associated islands. On cosmogonic and

TABLE 12.—*Average composition of plateau basalts (continents)*
(Calculated as water free)

Region	1 Deccan	2 Karroo, Rhodesia	3 Oregon	4 Spitzbergen-Franz Josef Ld.	5 Iceland	6 Faroe Island	7 Hebrides	8 Grand average
No. of analyses in average	8	6	2	3	4	2	5	30
SiO_2	50.67	51.12	46.34	49.54	48.03	49.12	46.98	49.25
Al_2O_3	13.30	13.69	13.87	14.17	13.61	14.19	15.56	14.05
Fe_2O_3	3.52	3.75	3.04	4.02	2.52	3.06	3.60	3.43
FeO	10.34	8.62	11.84	9.25	10.44	10.06	10.06	9.93
MgO	6.00	5.68	7.39	7.13	8.45	5.32	8.16	6.77
CaO	10.15	10.45	9.73	10.13	11.61	10.02	9.23	10.20
Na_2O	2.40	2.33	2.45	2.74	2.08	3.09	2.70	2.48
K_2O	.60	.81	.72	.58	.36	.76	.55	.62
TiO_2	2.52	2.94	3.58	1.76	2.56	3.82	2.68	2.71
P_2O_5	.31	.41	.85	.43	.16	.38	.22	.35
MnO	.19	.20	.19	.25	.18	.18	.26	.21
	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

geophysical grounds it is to be expected that chemically the substratum basalt of the continental sectors of the globe does not differ much from the substratum basalt of the oceanic sectors. On the other hand, the respective products of gas-fluxing and assimilation should differ considerably, because in the continental sectors these reactions are, by theory, largely concentrated in the sial, which seems to be lacking in most of the suboceanic crust. Here, then, we have a double test of the worth of the abyssolith hypothesis in explanation of the Hawaiian lavas.

To cover the first part of the test, let us compare the Hawaiian olivine basalt, assumed to be primary, with the material probably constituting the substratum under the continents. It is generally agreed that the plateau basalts were extruded so rapidly that their composition was little affected by contamination with crust rock or by the processes of differentiation. At the surface the flows of the plateaus have lost much volatile matter; otherwise they qualify well as representatives of the upper, basaltic, phase of the substratum in the continental sectors of the globe.

The best analyses of the plateau basalts are those of Washington (1922, p. 765). From his tables water-freed averages for regions in Asia, Africa, North America, and various parts of the North Atlantic field have been computed. The results are entered in columns 1 to 6 of Table 12. A similar average for the "plateau"

basalts of the Hebrides is given in column 7 (Bailey and Thomas, 1924, p. 157). The average of all of the 30 analyses is stated in column 8. The variation among the columns of figures is relatively small, and it seems that a fair estimate of the mean

TABLE 13.—*Comparison of continental plateau basalt with "primary olivine basalt" of Hawaiian Islands*
(Water free)

	1	2
	Continents (30 analyses)	Islands (23 analyses)
SiO_2	49.25	50.04
Al_2O_3	14.05	13.47
Fe_2O_3	3.43	2.65
FeO	9.93	9.05
MgO	6.77	7.79
CaO	10.20	10.49
Na_2O	2.48	2.43
K_2O	.62	.55
TiO_2	2.71	3.12
P_2O_5	.35	.29
MnO	.21	.12
	100.00	100.00
	NORMS	
Q	1.20	1.56
Or	3.36	3.34
Ab	20.95	20.43
An	25.60	24.19
Di	18.80	21.35
Hy	19.25	18.90
Mt	5.10	3.71
Il	5.15	5.93
Ap	.62	.62
Norm density	3.140	3.125

composition of the substratum (supposed to lack its actual content of water) under the continental surfaces is given by column 8.

That grand average is restated in Table 13, which also includes the preferred average analysis of the volatile-freed, "primary olivine basalt" of Hawaii, together with the respective norms and norm densities. Here again the differences in the pairs of oxides are not great, though the comparison suggests that the substratum may be slightly more femic under the mid-Pacific crust than under the continental crust. All these relations are consonant with good theory regarding the origin and thermal history of our planet.

We turn now to the second part of the double test. The Hawaiian trachyte

[2] The Hebrides average includes only five analyses, indicated by the symbols I, II, III, C, and D in the Mull memoir. Three other analyses listed on the same page were omitted because they do not refer to flows.

has been explained as the most salic residual liquid derived from the "primary olivine basalt" in an abyssolithic chamber, but increased in quantity by selective refusion and assimilation of the suboceanic crust, itself largely or wholly of basaltic composition. The trachyte of the deep-sea islands forms small masses—endogenous (crater-filling) domes, plugs, and overflows. In contrast, similar contamination of basaltic liquid with the relatively thick sial should ultimately give residual liquids richer in silica, though again in the form of endogenous domes, plugs, and overflows. Such bodies are numerous on the continents, and many of them are rhyolitic, not trachytic.

It should be emphasized that these rhyolitic bodies are products of volcanism of the central type and do not include the much more voluminous rhyolitic extrusives associated with batholithic invasion of the earth's crust. It is fairly obvious that neither gas-fluxing nor pure crystal-fractionation of primary basalt, nor their combination, can account for the batholithic colossi and the enormous floods of rhyolite typified by those of the Yellowstone Park. Elsewhere the writer (1938, p. 159–173) has published an explanation of batholiths and associated rhyolites in terms of major stoping and large-scale remelting of the sial. On the present occasion the batholithic problem will not be considered, and our thought will come back to the small rhyolitic bodies at central vents. Here it is easy to see that the gas-fluxing hypothesis has the merit of meeting a difficulty which has long troubled those petrologists who, with rightly placed confidence, ascribe the chemical diversity of the igneous rocks to the crystal-fractionation of highly femic magma. A given quantity of rhyolitic liquid, so differentiated and then collected at or near the earth's surface, should overlie more than 10 times its weight of complementary, ultra-femic material. Yet at many volcanic centers of the continents rhyolitic eruption is succeeded by basaltic eruption, not by that of ultra-femic material, either re-fused or in the form of rafted-together crystals. The actual sequence of lava types becomes intelligible if the concentrated rhyolite is chiefly the product of differential refusion of the high-lying sial and passes down into the little-modified basaltic melt which is capable of later eruption. There seems to be no objection to the view that the sial itself is essentially due to the primitive crystal-fractionation of the earth's (initially ultra-femic) mantle. Thus the theory of crystal fractionation becomes less troubled if it is combined with the idea of some (differential) assimilation in abyssolithic chambers. On this assumption much the greater part of the highly salic material was prepared long before these rhyolite-breeding volcanoes of the continents began their activity.

Since the "andesites" and their close allies in Hawaii have their analogues in the continental (central-vent) lavas intermediate between rhyolite and basalt, it is logical to suspect that the intermediate types of the continents are also products of mixing by two-phase convection.

ARE THERE TWO PRIMARY BASALTS IN CONTINENTAL SECTORS?

A general picture of petrogenesis at volcanic centers has been presented. In a thought-provoking paper Kennedy (1933, p. 239) has drawn one quite different, though he too relies on crystal fractionation as the essential cause of diversity

among the lavas, including those of Hawaii. He assumes the simultaneous existence of two primary basaltic melts, one of which he names "olivine basalt magma" and the other "tholeitic magma." The former is present in both oceanic and continental sectors (alkaline provinces) of the earth; the latter is largely confined to the continental sectors and is "represented by the great majority of world plateau basalts." He offers no suggestion as to the depths, volumes, and shapes of the primary bodies of liquid before they began to send offshoots through the earth's crust to the surface and thus does not commit himself on this fundamental, composite problem. His picture is further incomplete because of failure to show how crystal-fractionation, whether controlled by simple gravity or by "squeezing-out" of residual liquids, can get the long list of lava types from basaltic magmas. The possibilities of two-phase convection and refusion of crust-rock are not mentioned. His argument rests on the different potentialities of his two magmas as these slowly cool and pass, in contrasted ways, through their respective stages of crystallization. On this relatively simple premise Kennedy concludes that the olivine-basalt magma produced the "alkaline suite" of igneous rocks, successive members of which are listed as trachyandesite, trachyte, and phonolite. Similarly through crystal-fractionation of the tholeite magma he explains the "calc-alkaline suite"; representative of this suite are andesite (of orthodox definition) and rhyolite. The mineralogical and chemical divergence of the two suites is ascribed essentially to the fact that pyroxene, one of the early crystals to form in each primary melt, is lime-rich in the olivine-basalt case and lime-poor in the tholeite case. The separation of lime from the former melt means that

> "considerable amounts of lime are extracted in the crystallization and in consequence the residual liquid is impoverished in this constituent. The removal of lime in the pyroxene prevents it from combining with Al_2O_3 to form anorthite and the Al_2O_3 will be forced into combination with the alkalies, giving alkali-feldspar and, eventually, feldspathoids. In the tholeitic magma, on the other hand, lime is only removed in very small amount during the ferromagnesian crystallization and most of it is still available when feldspar begins to crystallize. In consequence, the Al_2O_3 will combine with most of the lime to give the anorthite molecule which forms, therefore, an important constituent even in relatively acid rock types. Thus although the original composition of the magma determines whether alkaline or calc-alkaline rocks will develop during differentiation, the separation of lime-rich or lime-poor pyroxene is the actual means by which the end is attained" (p. 253).

To the present writer Kennedy's theory is inadequate because it is based on two doubtful postulates: (1) that the essential processes in Nature can be read out of the laboratory crucible; (2) that the primary magmas actually have the compositions stated by Kennedy. That physico-chemical study of cooling silicate melts in the laboratory is utterly vital for the petrologist is clear as day, but it is equally manifest that no man-made furnace and crucible can ever imitate the complex conditions at an active volcanic vent. Differentiation within and beneath a volcanic throat is a matter of physics as well as of physical chemistry. Already we have caught glimpses of the complexities introduced by the combination; a full consideration of these is quite beyond the scope of this paper, which will, however, include a word about the second of Kennedy's postulates.

From a limited number of chemical analyses (some quite incomplete) of Mull and other Hebridean lavas Bailey and Thomas generalized and thought to account for the visible species of eruptive rock in that region by assuming them to be descend-

ants from an "olivine-basalt magma-type" and a second, "non-porphyritic central type" of basalt, also primary. These ideas were later generalized by Kennedy to cover the whole earth, without, however, going outside the Hebridean region for possible improvement of the two magma-types by collecting statistics from conti-

TABLE 14.—*Analyses relevant in discussion of Kennedy's two primary magmas*

	1	2	3	4	5	6
	Primary olivine basalt magma (Kennedy)	Primary tholeiite magma (Kennedy)	5 plateau basalts of Mull	5 nonporphyritic central lavas of Mull	Range of oxides in analyses of col. 4	9 parts of continental plateau basalt with 1 part of graphic granite
SiO_2	45	50	46.98	51.34	47.35–53.97	51.50
Al_2O_3	15	13	15.56	13.10	11.05–14.87	14.10
Fe_2O_3	13	13	3.60	4.07	2.73–5.87	3.11
FeO			10.06	9.61	6.32–13.08	9.03
MgO	8	5	8.16	5.36	2.58–6.84	6.15
CaO	9	10	9.23	9.96	6.36–12.89	9.31
Na_2O	2.5	2.8	2.70	2.97	2.36–3.48	2.47
K_2O	.5	1.2	0.55	1.13	0.54–2.27	1.34
TiO_2			2.68	1.87	0.94–3.25	2.47
P_2O_5			0.22	0.27	0.21–0.55	0.32
MnO			0.26	0.32	0.09–0.53	0.20
	93.0	95.0	100.00	100.00		100.00

nents and oceans. Thus in presenting his case he used the Bailey-Thomas estimates of composition for the Mull olivine basalt ("plateau") magma and the associated nonporphyritic central type of magma, which Kennedy renamed tholeitic magma. The two estimates are indicated by columns 1 and 2 of Table 14.

Bailey and Thomas (1924, p. 15) give eight analyses representing the more femic magma. Only five refer to flows—that is, to those bodies which are most likely to represent primary material. The average of all five is given in column 3, whose value for silica is distinctly greater than that in column 1.

These authors also give nine analyses (p. 17) that represent the nonporphyritic central or tholeitic magma-type. That they cannot be descriptive of any one magma is clear from column 5 of Table 14, which shows how great is the range in stated percentages for every one of the oxides. Excluding two incomplete analyses and three evidently inadmissible analyses of the "central" and "Talaidh" types, an average analysis for the less femic magma has been computed (also on a water-free basis) and entered in column 4 of Table 14. It closely resembles column 2.

A remark by Bailey (p. 4) suggests that it is *a priori* unwise to regard the Mull statistics as particularly helpful in the search for any primary magma to which petrogenesis throughout the continents as a whole could be referred. Bailey wrote that he was "doubtful whether the lavas [of Mull] still spared by erosion are not, in the main, the products of a central volcano, an idea always linked with the name of Professor Judd." We have seen how complex are the physical and physico-chemical

processes at long-lived central vents. If the rocks of Mull had been analyzed in greater number and with the accuracy and completeness illustrated in Washington's work, a better idea of the Mull olivine basalt would be obtained. Even with the few analyses available, the considerable degree of uniformity of composition characterizing the more mafic rocks of the island seems to argue for the primary nature of a well-averaged olivine basalt. Much less warranted is the choice of a second primary basalt among the species listed as belonging to the nonporphyritic central "type."

As an alternative the tholeitic magma of Kennedy may, without unduly violent speculation, be regarded as a mixture of average plateau basalt (as represented in all the continents) with a small amount of sialic rock. For example, let us assume that in a Mull central vent nine parts of this average basalt were mixed with one part of the abundant, low-melting material (quartz-feldspar eutectic) of the sial. The mixture would have the composition registered in column 6 of Table 14. Its similarity with column 4 of the same table is striking and gives color to the hypothesis that the tholeitic magma of the Hebrides was not purely primary material but a derivative from a syntectic.

The olivine basalt magma-type of Mull (column 1 of Table 14) seems inexplicable by any simple process of addition to, or subtraction from, the magma represented by plateau basalts the world over. The low silica and high alumina are particularly hard to account for. Here too one must ask whether the average composition and the actual range of composition reflect the results of vigorous mixing of phases by convection in the Hebridean abyssolith or abyssoliths.

In all this there is much pure speculation, but lack of definite answers to several questions cannot obscure the fact that the hypothesis of two separate primary basaltic melts under Mull and under the continents generally is, on thermal and structural grounds, far more difficult to accept than the hypothesis of one primary basaltic melt under the continents. However, this idea does not affect the conclusion already arrived at, that the substratum basalt under the islands of the deep ocean is somewhat more femic than the substratum under continental surfaces—perhaps slightly more femic than as illustrated in the average analysis of column 2 of Table 13.

CONCLUDING REMARKS

We look back. Once more many pages have illustrated the irony of Nature. She has endowed us with the urge to understand but hides much of the volcanic mechanism in the inaccessible depths. Yet no one who has had a good look at the turbulent, fountaining Halemaumau and at the results of its scientific study can accept defeatism. Geologists will continue their effort to see with the eyes of the mind how Vulcan and Pluto have worked. They will continue to speculate, and observations in Hawaii will long furnish the best controls over their imaginings. Yet even with this special help the attack on the double problem of volcanism and petrogenesis must remain speculative. In this respect the writer has been greatly daring; hence the welter of hypotheses described in the foregoing paper. He has made many extrapolations: from the findings of cosmogonist, petrographer, and

analytical chemist; from those of seismologists and geodesists; from the average thermal gradient in the rocks just below the earth's surface; from the visible phenomena at active craters, particularly those inside the Kilauea sink; and from the results of laboratory experiments on the development of phases in silicate melts and on the physical properties of rocks and rock glasses at different pressures and different temperatures.

The century-old idea of a world-circling, vitreous substratum as the source of the heat of erupted magma has been retained in spite of objections offered by seismologists and other experts in geophysics. That fundamental hypothesis has been considered in much detail elsewhere. Here more fully declared is emphasis on the effects wrought by the volatile matter, especially water, dissolved in the substratum. It is suggested that there the basaltic and peridotitic matter may contain dissolved water to the extent of something like 6 per cent by weight. That water is a flux, so powerful that the top of the substratum under Hawaii need not have a temperature higher than about 1300°. When the substratum melt is thrust up to the earth's surface, much of the water is exsolved. It is supposed to be concentrated, still in the free state, at the upper levels of the displaced magma, where the walls of the lava column are fluxed by the combined power of the gas and the heat of the invading magma.

By assuming the presence of so much volatile matter in the substratum we have theoretical explanation for: (1) the great lengths of basaltic flows, which remain at least in part liquid notwithstanding the rapid loss of heat as the lava runs; (2) the two-phase convection so characteristic of Halemaumau, Mokuaweoweo, Matavanu, and other basaltic vents, and the long duration of continuous activity at these vents; (3) the pipelike form, typical of Hawaiian and other central vents; (4) trachytic and other relatively salic lavas, developed largely by differential refusion of basic crust-rock; (5) rhyolitic plugs, endogenous domes, and flows at central vents on continents and their satellitic islands and shoals; (6) the great variety among lavas of intermediate composition—expected because of mixing of fractions by two-phase convection; and (7) at central vents, the common alternation of basaltic flows with highly salic flows.

The chemical diversity, both in the Hawaiian islands and throughout the world, is ascribed ultimately to the crystal-fractionation of primitive magma, but the liquids undergoing this fractionation include syntectics as well as basalt and primary peridotite. A new suggestion regarding the origin of the alkaline lavas of the volcanic piles built on the floor of the deep sea has been offered, but no attempt has been made to present the argument for limestone control in all its complex and controversial aspects.

Less troubled with speculation are certain by-products of this investigation. One of these is new and apparently cogent evidence that by the method proposed by Iddings we can get from the norm of an igneous rock a good value for the density which that rock would have if it were completely crystallized and quite pore-free. Using that method a rather close estimate of the change of density when a rock passes from a holocrystalline, pore-free state to a vitreous, pore-free state can be obtained. Such estimates are of manifest value in the theory of crystal-fraction-

ation as well as in theories about the nature of the earth's interior. Perhaps, too, the result of recalculating the average composition of the world's plateau basalt may be of use to petrologists.

But more significant than any quantitative statement of this paper is its main question: To what extent does juvenile volatile matter rival temperature in determining the state of the material concerned with volcanism and petrogenesis?

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ANORTHOSITES IN SOUTHEASTERN ONTARIO, CANADA

BY J. M. HARRISON

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ABSTRACT

Anorthosites, gabbros, and associated rocks in southeastern Ontario are described in detail. Field and laboratory evidence indicates that the anorthosites formed from a magma with a composition about that of gabbroic anorthosite. It is suggested that the more mafic masses are higher in the earth's crust than are the anorthosites. Rising volatile-rich fluids carrying iron, titanium, phosphorus, etc., and assimilation of limey sediments by the magma probably concentrated the dark minerals. The source of the magma may have been a layer in the earth's crust with a composition approaching that of bytownite anorthosite. Two granites are separated by the intrusion of the basic rocks.

INTRODUCTION

A geological investigation of Hinchinbrooke township and part of Bedford township, Frontenac County, southeastern Ontario, was undertaken during the field season of 1942. The area is about 30 miles north of Kingston and is part of the Grenville geological sub-province (Fig. 1).

Certain anorthosites in the area resembling the anorthosites of the Adirondack region were mapped and examined in detail in the hope of throwing some light on the problem of the formation of anorthosites. Two granites occur in the area, one earlier and one later than the anorthosite.

The area had not previously been investigated completely. W. G. Miller (1899) mentioned a few of the rock types seen in a reconnaissance of part of the area. Some deposits containing apatite and titaniferous magnetite that occur near Eagle Lake were commented on in 1890 (Rept. Roy. Comm., p. 167) and were described by Spence (1920). Cousineau (1940) further described these deposits in some detail.

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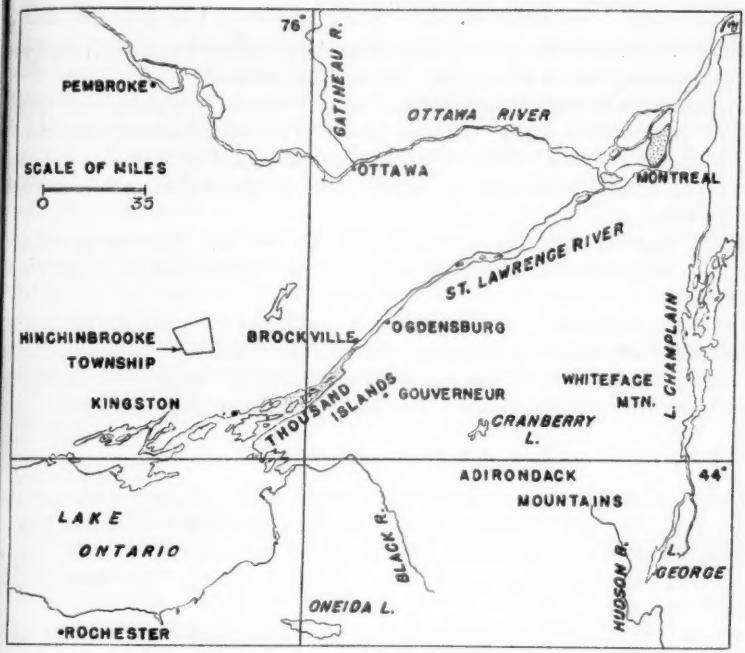

FIGURE 1.—*Index map showing location of Hinchinbrooke township*

GENERAL GEOLOGY

GENERAL STATEMENT

All consolidated rocks in the area mapped are believed to be pre-Cambrian. The geologic succession is as follows:

QUATERNARY — Unconsolidated gravels, sands, and soils

.....Great Unconformity.....

PRE-CAMBRIAN — Trap dikes

Intrusive Contact

Pegmatite and felsite dikes

Alaskitic granite, generally pink and massive, some syenite at Wilkinson

Intrusive Contact

Pegmatite dikes

Anorthosite, gabbroic anorthosite, and diorite.

Intrusive Contact

Pegmatite dikes

Gneissic granites—Hinchinbrooke granite and some smaller masses of gneissic granite and syenite.

Intrusive Contact

Grenville series—limestones, paragneisses, quartzites, and possibly some volcanics.

GRENVILLE SERIES

The Grenville series consists of much altered sediments—limestones, quartzites, and paragneisses—and a minor group that may be volcanic.

The limestone, interbedded with sedimentary gneisses and quartzites, is impure and coarsely recrystallized, but in most of it non-carbonate constituents of metamorphic origin are rather sparse. Graphite is the most widespread of these, but pyroxene, amphibole, apatite, muscovite, phlogopite, talc, and serpentine are common and, locally, abundant.

Several types of bedded paragneiss occur in the district. Some are almost pure quartzite, others are feldspathic, and some are rich in garnet, biotite, amphibole, or pyroxene.

In the northwest section of Hinchinbrooke township a green gneiss with irregular light-colored layers may be of volcanic origin. Where jointing has provided a surface parallel to the strike of the schistosity but about 30° steeper than the dip, the light layers are distinctly elliptical in outline, suggesting original pillow lava.

HINCHINBROOKE GRANITE

Introduction.—This granite, the most widespread of the gneissic granites, underlies most of the south half of Hinchinbrooke township, the southwest portion of Bedford township, and occurs as far north as the southwest corner of Oso township. Other smaller masses were noted but were not studied in detail. The Hinchinbrooke granite is pre-anorthosite.

Wilson (1930) considered that there are two ages of batholithic granite in the Westport area, which adjoins the Hinchinbrooke area on the east, but he noted "... so far as known, no rocks either intrusive or sedimentary occur anywhere in the Grenville sub-province that separates the two groups of batholithic rocks from one another." He found evidence, however, of two ages of granite near Montreal, Quebec. He described (1925, p. 396–399) a gneissic granite cut by diabase dikes and intruded by two small stocks of granite which do not intrude the overlying Potsdam sandstone.

Miller and Knight (1913, p. 1–58) first suggested the existence of two granites in southeastern Ontario, largely on lithologic grounds. Two granites were distinguished by Baker in the Kingston district (1916, p. 12–15), also largely on the basis of lithology. Wright, however, found no positive evidence for two granites (1923, p. 32–33). He traced the granite of the Brockville area west to Baker's Algoman (younger) granite and south to granite in the Thousand Islands district called Laurentian (older) by Cushing (1910, p. 36–38).

Osborne (1936; 1938) mentioned a pre-anorthosite granite in the Grenville north of the St. Lawrence River in Quebec but presented no evidence for so classifying it. Faessler (1943, p. 132) reported a pre-anorthosite granite in the Grenville of Papineau County, Quebec, but left detailed description for a later paper.

Buddington (1939, p. 197–201) critically reviewed the evidence for a granite preceding intrusion of the anorthosites in the Adirondack region of the Grenville subprovince and concluded that it was not satisfactory.

Petrography.—The typical Hinchinbrooke granite contains a large percentage of

quartz, though locally it is rather sparse. Amphibole, biotite, and pyroxene are the main accessories. Magnetite is usually present as a minor constituent. Quartz blebs and grains are conspicuously elongated and weather as little ridges so that the foliation is distinct even where very small amounts of mafic minerals occur.

Thin sections of the granite show it to consist of quartz, plagioclase, and potash feldspar with lesser amounts of amphibole, biotite, pyroxene, and magnetite. Minor accessories are apatite, sphene, and rare zircon. The dark silicates are largely altered to intergrowths of chlorite, carbonate, and epidote, but the feldspar, except locally, is fresh with only minor amounts of cloudy alteration products. Feldspar and quartz invariably show strain extinction and have interlocking borders. In some thin sections orthoclase occurs as antiperthite, but in most it cuts across oligoclase and quartz grains.

An average of Rosiwal determinations from eight thin sections of granite from widely separated localities gave the following composition:

	per cent		*per cent*
Oligoclase (An_{30})	44	Amphibole	3
Orthoclase	12	Biotite	4
Quartz	33	Magnetite	2
Pyroxene	2		

These Rosiwal determinations indicate a soda granite, which is very similar in lithology and mineralogy to the "older" gneissic granite of the Westport area (Wilson, 1930).

Some pegmatite dikes associated with the granite are very similar to it except for their larger grain size and are very probably related to the granite.

Relations to the Grenville series.—Inclusions of altered Grenville paragneisses are common in the Hinchinbrooke granite, especially near the margins. Some inclusions are ghostlike remnants with indefinite boundaries, but others stand out sharply from the enclosing granite. In some parts no inclusions were found for distances of 2 or 3 miles across the strike of the foliation. It is likely that the composition of the granite has been modified by incorporation of the Grenville rocks, but the presence of pegmatite dikes and constant character of the granite over large areas makes it unlikely that it is a sodic Grenville paragneiss with characteristics of a granite.

ANORTHOSITES AND RELATED ROCKS

The following brief resume of the relations of the anorthositic rocks to the Hinchinbrooke granite indicates that the anorthosite postdates the gneissic granite. The evidence is discussed in the section dealing with detailed descriptions of the basic rocks.

In general, the basic masses are concordant with the structure of the enclosing rocks. They are coarse-grained in the central portions and finer-grained toward the margins, whether the enclosing rock be Grenville gneisses or Hinchinbrooke granite. In a few places near the margins of some of the intrusives, the granite is cut by dikes identical with the finer-grained margins of the basic rocks. Further, the granite is cut by dikes such as would be expected as differentiates from the anorthositic rocks. Finally, titaniferous magnetite, characteristic of the basic rocks, occurs as irregular stringers and grains in granite near its contact with the basic rocks.

ALASKITE

The main occurrence of alaskite is along the east boundary of Hinchinbrooke township, from the village of Tichborne south to St. Andrew Lakes. The alaskite is fine- to medium-grained, pink to dark red, with blocky phenocrysts of microcline from one-eighth to three-eighths of an inch across. Anhedral quartz grains of about the same dimensions are locally present. Biotite is the dominant accessory, though muscovite is occasionally present.

Under the microscope typical alaskite appears massive, medium-grained, and porphyritic (Pl. 1, fig. 1). Some of the phenocrysts of microcline contain rounded blebs of quartz in parallel orientation, and some contain two groups of blebs. Those of each group extinguish simultaneously but at a different position from those of the other group. Adjacent grains of plagioclase and microcline contain optically parallel quartz grains (Pl. 1, fig. 2). No satisfactory explanation for this phenomenon was found.

An average of four Rosiwal analyses on thin sections from the alaskite gave the following composition:

	per cent		per cent
Orthoclase	35	Mica	3
Albite (An_7)	33	Magnetite	1
Quartz	28		

This granite is very similar to certain phases of Wilsons "younger" granite of the Westport area (1930).

Dikes and apophyses of alaskite cut all rocks thus far mentioned and in some places have produced a pink color in the anorthositic rocks, especially near the alaskite. Contacts of alaskite with the Hinchinbrooke granite and the Grenville gneisses are similar.

TRAP DIKES

A few fine-grained, massive, basic dikes cut all rocks in the area. They are slightly diabasic and consist mainly of andesine (An_{40}), amphibole, and some magnetite.

ANORTHOSITE, GABBROIC ANORTHOSITE, AND RELATED ROCKS

SUMMARY OF ANORTHOSITE PROBLEMS

The theories of origin for anorthosite fall into two main groups. In one the anorthosite is considered never to have been molten as such, and the other assumes a magma of essentially anorthositic composition.

One difficulty in assuming an anorthositic magma is the high temperature generally considered necessary to keep it molten. According to experimental work by Bowen (1928, p. 189), a melt consisting of labradorite with 20 per cent diopside has a melting point of 1350°C. Some years earlier (1917) Bowen presented his arguments against the existence of a magma with the composition of anorthosite, among which were the high temperature necessary for a molten anorthosite, paucity of anorthosite dikes, and universal protoclastic structure. He suggested that anorthosites were formed by settling of labradorite crystals, above a layer of mafic minerals, from a crystallizing

gabbroid magma. The granite-syenites, universally associated with anorthosite, he regarded as the liquids which crystallized after the separation of the early minerals.

Balk (1930; 1931) modified Bowen's theory somewhat by suggesting that the labradorite crystals were accumulated by "pressure flowage" and that the granite-syenite represents the "mother liquor" driven upward and outward from the labradorite crystals. This would explain the occurrence of labradorite phenocrysts in the syenite.

Largely on the basis of field data, many have objected to the theory that anorthosites are merely crystal accumulations. Among authorities who have presented evidence for an anorthosite magma are Alling (1932), Buddington (1939), Cushing (1917), Fowler (1930), Kemp (1921), Mawdsley (1927), and W. J. Miller (1931). Practically all these writers believe that the granite-syenite series is distinctly later than the anorthosites with which they are associated.

Grout and Longley (1935) found that the anorthosite facies of the Duluth gabbro are closely related to red granite containing xenocrysts of labradorite with which the granite has reacted. They believe, however, that the anorthosite was solid when intruded by the granite. Thus, mechanical forces can, and do, isolate phenocrysts (xenocrysts) from anorthosite.

The most comprehensive memoir published on the problem of anorthosites in the Adirondack region of the Grenville subprovince is that of Buddington (1939). He presents much evidence for an anorthosite magma which, he suggests, had a composition about that shown in Table 2, E, and that the mafic minerals were removed in part by segregation, in part by sinking, and in part by rising. The source of this magma is believed to have been a layer in the earth's crust which may have had a composition approaching that of bytownite anorthosite. The concept is carried further in a later paper (Buddington, 1943) in which he suggests that a shell of bytownite anorthosite may exist about 25–30 kilometers below the earth's surface. This shell may have been formed by primordial differentiation much as it has been shown to occur in large sills of gabbroic rock now exposed. Since selective melting of this layer is suggested as the source of the labradorite anorthosite, the composition of the shell must change to one containing only the constituents of higher melting point and would not again give rise to anorthosite from that locality. Further, geoisotherms were probably higher in the early part of the earth's history, and the layer could be tapped more easily. These conditions would cause the general restriction of anorthosites to the pre-Cambrian as pointed out by Daly (1933, p. 412).

ANORTHOSITE BODIES IN AREA

The terms anorthosite, gabbroic anorthosite, and anorthositic gabbro have been defined by Buddington (1939, p. 19) as rocks containing respectively 0–10, 10–22½, and 22½–45 per cent of mafic minerals. However, rocks in this area which are gabbroic anorthosite according to this definition have been referred to as gabbros for more than 40 years (Miller, 1899), and to avoid excessive use of a rather cumbersome name the terms are used synonymously in this paper.

One body of anorthosite and six of gabbroic anorthosite, with anorthosite phases, were mapped (Fig. 2).

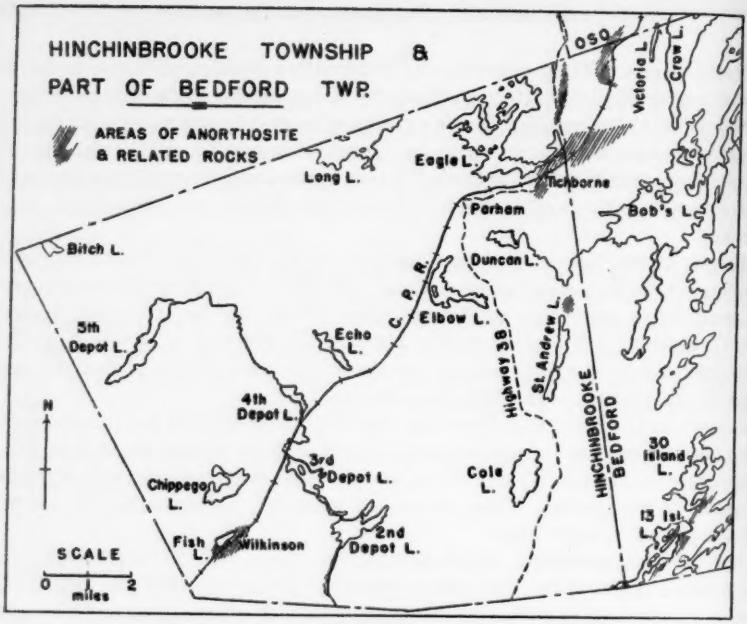

FIGURE 2.—*Map showing location of the anorthositic masses* Hinchinbrooke-Bedford area.

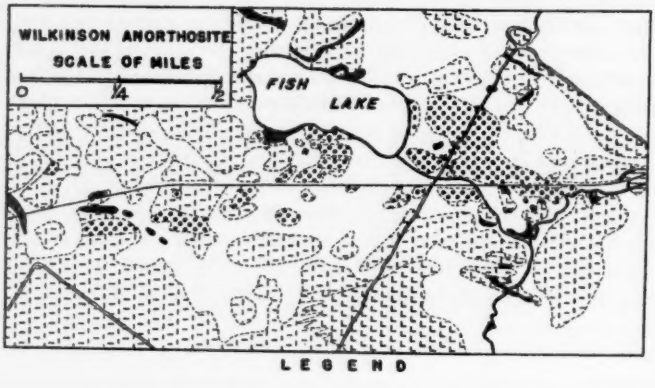

FIGURE 3.—*Wilkinson anorthosite*

WILKINSON ANORTHOSITE: Anorthosite outcrops in the immediate vicinity of Wilkinson station, in the southwest corner of Hinchinbrooke township (Fig. 3).

TICHBORNE GABBROIC ANORTHOSITE: This is the largest body examined, and it was mapped more completely than the other, similar, bodies. From a point about three-quarters of a mile southwest of the village of Tichborne it extends in a flat U along and near the south shore of Eagle Lake and eastward for about 1½ miles (Fig. 4).

BLESSINGTON GABBROIC ANORTHOSITE: This rather small mass surrounds the old Blessington (or Eagle Lake) iron and phosphate mines. From a point near the north end of the east arm of Eagle Lake it extends north to the southwest corner of Oso township. It is similar to the Tichborne mass (Fig. 4).

VICTORIA GABBROIC ANORTHOSITE: This mass occurs west of Victoria Lake in the northeast part of the map area; its northern limits were not determined. Some portions of it are extremely coarse-grained, but otherwise it is similar to the other units (Fig. 4).

ST. ANDREW GABBROIC ANORTHOSITE: This is a small but very coarse-grained unit exposed just north of St. Andrew Lakes. Presumably a large proportion of this mass has been destroyed by alaskite for no fine-grained border zone is exposed.

ISLAND LAKES GABBROIC ANORTHOSITE: This mass, in southwest Bedford township, was not completely outlined. It is similar in mineralogy and lithology to the Tichborne mass.

OCONTO GABBROIC ANORTHOSITE: The south end of this body occurs in the southwest corner of Oso township. Its other limits are unknown.

In his brief examinations W. G. Miller (1899) included all the basic rocks he saw under the name "Parham gabbro" because of the occurrence of gabbro at what was then Parham station. Parham station, however, is now Tichborne, and, as it is also known that the basic rocks occur as several separate bodies, the name "Parham" has been dropped, and each unit has been given a name. The Wilkinson anorthosite has not previously been described.

PETROGRAPHY OF ANORTHOSITE AND RELATED ROCKS IN THE HINCHINBROOKE AREA

GENERAL STRUCTURE

All the gabbroic anorthosites of this area are so alike in general character that inferences drawn from one apply to all. The Wilkinson anorthosite, however, has some distinctive features.

Few satisfactory data were obtained regarding the internal structures of the masses. Marked foliation is confined chiefly to the margins where platy elements indicate a general conformity with the foliation of enclosing older rocks.

An eastward dip characterizes the Blessington gabbro particularly. The Tichborne and Victoria masses dip generally eastward, but local reversal of dip is quite common. Available data indicate that the linear elements in the three gabbros plunge eastward and northward at angles usually less than 60°.

Block structure, as described by Balk (1930, p. 297) as characteristic of many Adirondack anorthosites, was not recognized in any of these masses. Jointing is

FIGURE 4.—*Tichborne, Blessington, and Victoria gabbroic anorthosites*

locally well developed in the gabbros, and the joint surfaces are commonly coated with prismatic scapolite. At some localities in the Wilkinson body rather light-gray very coarse anorthosite is cut by narrow dikelets of finer-grained, light-pink material that consists of andesine (An_{32}), with minor amounts of quartz, antiperthite, microcline, magnetite, biotite, and a very little tourmaline. Some of these appear to fill joints in the anorthosite.

WILKINSON ANORTHOSITE

Megascopic description.—The Wilkinson anorthosite consists typically of coarse crystals of plagioclase with fine-grained, dark, interstitial material. Locally it weathers dark gray, but more commonly it is very pale gray. The size of the plagioclase crystals varies in some measure as the distance from the margins, the largest crystals occurring in the central part where platy to tabular individuals 2 inches or more long and more than an inch wide are common.

Microscopic examination.—Examination of thin sections was confined largely to the finer-grained phases of the anorthosite. For all anorthositic rocks the composition of the plagioclase was determined by the oil-immersion method on cleavage flakes. It may be noted here that, for part of the curve, the values given by Tsuboi (1923, p. 122) for N'_p on cleavage flakes of plagioclase are less than those given by Winchell (1933, p. 318) for N_p of plagioclase of the same composition but are in fair proportional agreement with the values for N_p given by Chudoba (1933, p. 21).

Most of the large crystals of plagioclase have bent twinning striae, show strain extinction, and are rimmed with finer, granular plagioclase. A blocky texture is characteristic of the interior of the mass, whereas diabasic texture is prominent in some of the more gabbroic border phases. Protoclastic (cataclastic?) structure is characteristic of parts of all facies.

The plagioclase of the anorthosite varies from An_{45} to An_{53}, averaging An_{47}. Potash feldspar occurs sparingly in the main anorthosite, partly as minute inclusions in plagioclase, partly as narrow, branching stringers that cut other minerals, and, rarely, as interstitial grains. Some specimens taken at or near the contact between anorthosite and younger granite contain quartz. In one section irregular, isolated, but optically parallel grains of quartz replace diversely oriented grains of plagioclase (Pl. 1, fig. 3).

Pyroxene is the most common dark mineral. Both augite and hypersthene are common in the more gabbroic portions. Pyroxenes are almost invariably rimmed with uralitic amphibole and biotite. Titaniferous magnetite and apatite are present in all facies, but are more abundant in the gabbroic types. Sphene commonly occurs in the more gabbroic varieties, particularly near inclusions of limestone.

There appears to be no variation in the composition of the minerals toward the borders of the mass, but the dark minerals become more abundant. At most places the gabbroic facies grade abruptly outward to a rock containing much potash feldspar and a less calcic plagioclase. No sharp contact was seen, but at some places the gradation is rather abrupt. At one place the transition from anorthosite to syenite takes place within 100 feet and with no intervening gabbroic phase. As the syenite is approached fine-grained, pink, syenitic, interstitial material increases until finally the

rock is typical syenite with very sparse, scattered phenocrysts or xenocrysts of dark plagioclase. Narrow stringers of orthoclase occur in anorthosite near the contact with the syenite (Pl. 1, fig. 4).

The rock bordering the anorthosite is generally syenite or diorite with oligoclase (An_{18-28}). The phenocrysts of plagioclase in the syenodiorite vary from An_{18} to An_{36}. Other phenocrysts examined consist of microperthite. The composition of these crystals bears no apparent relation to the distance from anorthosite though they are more abundant near it. Some are well formed, but others are noticeably rounded.

Chemical composition.—Two chemical analyses of the Wilkinson anorthosite are shown in Table 1 (A and C) with an analysis of corundum-bearing bytownite anorthosite about 25 miles to the northeast. The last is the only other analysis of anorthosite from this general area and is of interest because it is of bytownite anorthosite.

The analyses of Table 1 are shown recast to minerals in the same table. The recasts of analyses A and C are to approximate modes. Data for the minerals of specimen B are lacking except that the composition of the plagioclase is An_{30} (W. G. Miller, 1899, p. 227).

The specimens for analysis A contained stringers and dikelets of less calcic anorthosite as described under Wilkinson Anorthosite. The main anorthosite must once have been typical labradorite anorthosite containing little besides titaniferous magnetite as accessory. The percentage of titaniferous magnetite is greater than was estimated from microscopic examination of the rock unless the microlites in the grains of plagioclase are iron oxides. Similarly, the presence of 5.34 per cent orthoclase indicates that some potash feldspar must be in solid solution in the plagioclase. Some corundum appears in the recast of analysis C, but no corundum is seen in the rock. Hence, the alumina probably occurs in the pyroxenes and amphiboles.

GABBROIC ANORTHOSITES

Description.—The gabbroic anorthosites vary a great deal in color. The more anorthositic phases, which are invariably coarse-grained, weather light gray, becoming darker with increasing content of mafic minerals. Medium-grained varieties vary from light to dark gray to almost black, depending on the mineral constitution. The finer-grained border phases typically weather light to dark brown and are medium to dark brown on fresh surfaces.

Some parts of the gabbros weather distinctly pink, particularly the medium- to fine-grained outer portions. However, some pink zones are extremely coarse-grained. Even in the portions in which the pink color is most pronounced, laths of purplish-brown plagioclase are prominent. The intensity of the pink color in the Tichborne gabbro increases irregularly but progressively to the east. The Victoria and Blessington gabbros also contain pink phases, and at many places in all these masses the pink color appears to be related to intrusions of pink alaskite.

All these gabbros have some very coarse-grained phases, usually in the central portions of the masses. Laths of dark-purplish plagioclase are as much as 3 inches long. In general, where the full sequence of grain sizes is developed in these gabbros, a dense, fine-grained rock forms the borders.

The Tichborne gabbroic anorthosite was selected as representative of this type of rock, and detailed laboratory examination was confined to specimens from it taken at regular intervals across the intrusive and from other places considered likely to furnish critical evidence. Supplementary specimens were collected from other similar intrusives. Four specimens from the Tichborne intrusive were analyzed chemically.

The coarser varieties of the gabbroic anorthosites are similar to the Wilkinson anorthosite but usually contain more dark minerals. In approximate order of decreasing abundance the primary minerals are plagioclase, pyroxene, amphibole, biotite, titaniferous magnetite, potash feldspar, apatite, scapolite, quartz, sphene, pyrite, and very rare zircons. Secondary alteration products include the cloudy products of the feldspars, carbonate, chlorite (and serpentine?), zoisite or clinozoisite, and limonite.

Feldspars.—All compositions of plagioclase here recorded were determined by oil-immersion method on cleavage fragments. Methods ordinarily used in thin section were not at all reliable. In some cases determinations made on grains of plagioclase, oriented normal to X or Z, indicate that the composition of individuals in the same thin section differ as much as 10–15 per cent in anorthite content. This variation is probably an effect of strain. The composition of the plagioclase based on such data, though not strictly accurate, will be sufficiently so for a general study. Examination of a number of thin sections with the universal stage reveals that pericline twinning is at least as abundant as albite twinning.

In the Tichborne gabbroic anorthosite the composition of the plagioclase varies from An_{51} near the west end to An_{18} near the east end. Usually the coarser the rock, the more calcic the plagioclase, though some very coarse-grained phases contain oligoclase.

The compositions of plagioclase in 55 specimens from the Tichborne intrusive were plotted on an outline of the mass. Lines were drawn enclosing areas of plagioclase of certain composition (Fig. 5). The anorthite content of the plagioclase decreases fairly regularly to the east and outward from the center of the intrusive. The south half of the broad east portion is less uniform, and "lows" in anorthite content may be ascribed to introduction of soda and silica from near-by alaskite.

Thin sections were etched with hydrofluoric acid and stained with sodium cobaltinitrate to determine the potash feldspar content (Keith, 1939). Orthoclase was found in many thin sections from the more basic portions, but generally not more than a few small grains occur in each section. However, one or two thin sections containing calcic andesine have as much as 5–10 per cent orthoclase. The proportion of orthoclase is considerably greater at the east end of the Tichborne intrusive, where, in a few places, it forms as much as 30 per cent of the rock (Fig. 5).

Orthoclase occurs in a variety of habits—in the more basic portions of the gabbros as small irregular grains interstitial to plagioclase and concentrated near the dark silicates, in other places as small stringers in perthite and antiperthite. In the east end of the Tichborne mass individual grains are numerous. Much of the perthitic rock appears to be the replacement variety (Pl. 1, fig. 5).

Orthoclase in the Tichborne gabbro is, in general, related to intrusive alaskite. At a number of places zones of friable pink syenite within the gabbro are cut by numerous

dikes of pink alaskite. At the west end of the intrusive, no alaskite is exposed where an abnormal amount of orthoclase is present. It is likely that the effects are due to unexposed alaskite. Toward the east end of the mass the high percentage of orthoclase is probably due to the presence of numerous small dikes of alaskite. Here

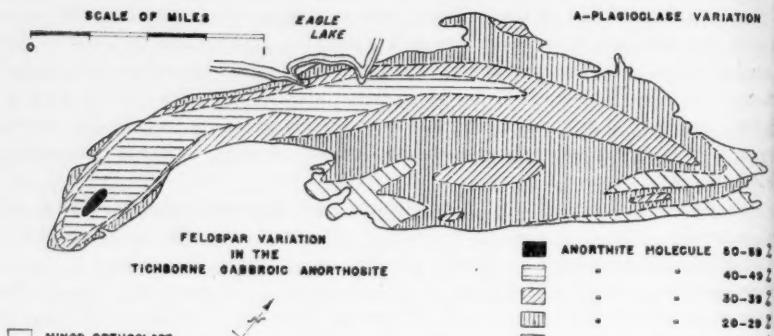

FIGURE 5.—*Feldspar variation in the Tichborne gabbroic anorthosite*

the effects of the alaskite are more pronounced, as the plagioclase is abnormally sodic. This rock resembles somewhat the laurvikites and nordmarkites tabulated by Kemp and Alling (1925, p. 46–47), except that it contains considerable biotite, locally lacks pyroxene, and is distinctly pink rather than green-gray.

Accessory minerals.—The chief accessory in most phases of the gabbros is pyroxene. Two types are present. Where only one variety occurs faint pleochroism indicates a relatively high percentage of hypersthene and/or hedenbergite molecules. Hypersthene is quite common and occurs in specimens that contain pale to colorless normal augite.

In nearly all thin sections examined the pyroxene is partly altered to amphibole and/or biotite. Alteration of pyroxene to amphibole is generally more complete in the pinker portions, and in some such thin sections no pyroxene is left. However, in some thin sections from pinker parts of the Tichborne mass, pyroxene is the main accessory, though partly altered to biotite.

Amphibole is clearly uralitic in most of the sections, but in some there is no evidence that it is not an original constituent. In some places amphibole appears in stringer-like arrangement between grains of feldspar.

The composition of the amphibole is variable, and many sections contain two varieties. In such cases pale blue-green, fibrous, actinolitic amphibole is apparently partly altered to a deep blue-green, more blocky hornblende. Deep greenish-brown amphibole in a few thin sections is probably "basaltic" hornblende.

Biotite occurs in nearly all specimens of gabbro. Some of it is a poikilitic alteration product of pyroxene and amphibole. In many sections it occurs as stringers that cut across feldspar, and it commonly rims titaniferous magnetite.

Magnetite is present in all thin sections examined and has a variety of habits. The larger grains are intergrown with ilmenite, but some of the smaller grains are probably free of titanium. Small amounts of magnetite occur as isolated blebs in large grains of plagioclase. Minute rod- and beadlike inclusions of opaque material, present in practically all grains of plagioclase, are too small for certain identification, but are probably magnetite. Analysis A, Table 1 shows more magnetite than appears in the rock analyzed unless the microlites are iron oxide.

Magnetite occurs as blades and grains in pyroxene, as an alteration product formed during uralitization, and as lamellae along cleavages of, and enclosed by, biotite that formed from amphibole. These three types are probably the result of breaking down of original iron-rich minerals to those containing less iron.

Finally magnetite occurs in stringers that cut all minerals so far mentioned. This late magnetite occurs in larger grains than the other varieties and is the most abundant type in the gabbro. It is invariably titaniferous. In some instances it is so intimately intergrown with biotite as to suggest that they were contemporaneous, though elsewhere stringers of magnetite cut across blades of biotite.

Small, irregular grains of pyrite, present in many thin sections, are always near late titaniferous magnetite and are commonly intergrown with it.

Apatite occurs as small clumps or as irregular stringers that cut the silicates. Only magnetite encroaches on apatite; the two are common together and are probably nearly contemporaneous.

Sphene is fairly common in the east end of the Tichborne mass and in the more sodic portions of the other masses. In most places it coats late titaniferous magnetite or is intergrown with apatite. In a few places it apparently follows cleavages in biotite.

Quartz is present in a large number of thin sections but rarely exceeds 5 per cent of the rock, and then only where alaskite intrudes the gabbro. In the more basic portions quartz, if present, occurs usually as small particles included in poikilitic amphibole. It occurs in greater amounts in the pinker portions as small irregular grains scattered through feldspar.

Scapolite is fairly common in the gabbros and is most abundant in the coarser-grained phases. It does not appear to have any particular relations to the margins of the intrusives, but, at one place on the west side of the Tichborne mass, the gabbro is bordered by rock that consists almost entirely of granular scapolite and amphibole. This rock grades into limestone by decrease in silicate constituents and has obviously been formed from limestone by action of the gabbro.

Within the gabbros scapolite occurs typically as ragged, branching stringers that transect numerous grains of plagioclase. The striking feature of these stringers is that the scapolite usually is optically continuous even though it has replaced parts

TABLE 1.—*Chemical and modal analyses of Wilkinson anorthosite*

In weight per cent

	A	B	C
SiO_2	55.29	47.32	45.12
Al_2O_3	26.02	30.36	17.60
Fe_2O_3	1.60	1.35	4.77
FeO	1.30	1.55	7.29
MgO	0.28	2.44	6.05
CaO	8.63	15.45	11.71
Na_2O	5.71	1.88	2.66
K_2O	1.03	0.66	0.49
H_2O (+)	0.13	0.10	0.53
H_2O (−)	0.10	n.d.	0.10
CO_2	0.07	0.58	0.56
TiO_2	0.52	n.d.	2.24
P_2O_5	0.03	n.d.	0.08
MnO	n.d.	n.d.	0.08
FeS_2	0.05	n.d.	0.79
	100.76	101.69	100.07

MODES

	A	B	C
Quartz	—	—	—
Orthoclase	5.34	1.95	—
Albite	48.27	15.20	20.96
Anorthite	43.17	64.43	25.07
Hypersthene	—	1.46	6.25
Diopside	—	4.00	18.75
Actinolite	—	3.15	3.39
Hornblende	—	2.24	3.10
Biotite	1.14	2.15	3.66
Sericite	—	0.79	0.79
Scapolite	—	—	1.25
Calcite	0.16	1.32	1.20
Sphene	—	—	0.59
Magnetite	2.32	1.96	6.89
Ilmenite	0.99	—	3.79
Corundum	—	1.37	1.64
Pyrite	0.05	—	0.79
Apatite	0.06	—	0.19
Kaolin	—	1.94	1.54
	101.50	101.96	99.85
Plagioclase	$Ab_{55}An_{45}$*	$Ab_{20}An_{80}$	$Ab_{47}An_{53}$
Mafics	4.50	14.96	47.21

* Plagioclase of 2 compositions present: (a) $Ab_{48}An_{52}$—original plagioclase of anorthosite; (b) $Ab_{55}An_{45}$—introduced plagioclase.

A—Composite of three specimens from central portion of Wilkinson anorthosite. Analyst, J. M. Harrison.

B—Anorthosite of South Sherbrooke Township, Lanark County, about 25 miles northeast of Wilkinson, from W. G. Miller (1899, p. 227). Analyst, Wm. Lawson.

C—Gabbroic rock near northeast border of Wilkinson anorthosite. Analyst, W. F. Green. Analysis supplied through the courtesy of Dr. W. D. Harding, Ontario Department of Mines.

of several grains of plagioclase of diverse orientation (Pl. 1, fig. 6). In some specimens the marialite-meionite (Na–Ca) ratio in the scapolite and the albite-anorthite ratio of the plagioclase are approximately the same, but in others the scapolite is considerably more calcic than the plagioclase. Scapolite cuts all other silicates except sphene and was probably formed by reaction of the gabbro with limestone.

Alteration.—All primary minerals except sphene are somewhat altered. In some instances the dark silicates of a specimen are much altered, whereas the light silicates appear quite fresh.

Feldspars and scapolite commonly contain cloudy alteration products. Sericite (or paragonite) is locally abundant, and small amounts of carbonate may occasionally be recognized. Rarely, small grains of zoisite or clinozoisite occur in altered feldspar.

Pyroxenes are commonly altered in some degree to mixtures of chlorite, carbonate, and magnetite and are generally more altered than amphiboles.

Biotite ranges from a fresh, deep-brown type through rather green, but highly birefringent material, to deep-green, almost isotropic chlorite. In a few instances some carbonate appears to have formed during the alteration.

CHEMICAL COMPOSITION OF THE TICHBORNE GABBRO

Four samples from the Tichborne intrusive were analyzed (Table 2), at which time it was not realized how great an effect the alaskite had had on some portions of the mass, and specimens containing introduced potash were included in the composite sample A. Hence the analysis shows more potash, and probably silica, than the gabbro originally contained. It is like a gabbro in amount of silica, total iron, titania, and phosphorus pentoxide, but more like a diorite in percentage of alumina and alkalis. However, if much of the potash and some of the soda are introduced, the rock originally was gabbroic. Specimen B is a fairly typical gabbroic anorthosite. Except for the lower content of lime it is very similar to gabbroic anorthosite from the Adirondacks (Table 2, E). Specimen C was taken 150 to 200 feet nearer the contact of the gabbro and is clearly more gabbroic. Specimen D was taken from the pink, eastern portion of the mass. Except for lower content of silica it is very similar to an augite-hypersthene syenite from the Adirondacks (Table 2, F) and contains extraordinary amounts of iron, titania, and phosphorus pentoxide.

Proportions of minerals obtained in the recasts (Table 2) are in fair agreement with amounts seen in thin section. The striking feature of recast A is the large proportion of titaniferous magnetite for the intrusive as a whole. However, some zones in the mass contain 25–30 per cent by volume of metallic minerals.

The FeO:MgO ratio in the silicates of the Tichborne is fairly constant (Table 3). It is lower for the intrusive as a whole than it is for the syenitic end and the more anorthositic portions. The facies near the west border has a ratio about the same as that of the whole intrusive. Therefore, in the more silicic and feldspathic portions of the mass there may be a concentration of FeO with respect to MgO.

RELATION OF COMPOSITIONAL VARIATIONS IN THE GABBROS TO THEIR STRUCTURE

The Blessington gabbro, as shown by foliation, dips consistently about 60° E. The west, or lower, part is rather fine-grained and consists mainly of plagioclase (An_{25}–An_{35}) with accessory pyroxene, amphibole, titaniferous magnetite, and apatite.

TABLE 2.—*Analyses of rocks from the Tichborne intrusive*

	A	B	C	D	E	F
SiO_2	49.50	52.00	48.72	55.20	53.40	58.26
Al_2O_3	17.63	24.44	19.91	15.93	23.96	15.51
Fe_2O_3	5.34	1.36	3.86	3.57	0.91	2.02
FeO	6.61	3.80	6.01	5.01	3.02	6.64
MgO	3.82	1.66	3.81	2.10	1.88	1.77
CaO	7.15	7.60	9.14	4.60	9.85	4.29
Na_2O	4.52	4.54	3.76	4.12	4.17	4.04
K_2O	1.80	1.13	0.89	3.76	0.80	4.46
H_2O (−)	0.09	0.04*	0.10	0.04*	0.04	0.07
H_2O (+)	0.38	0.40*	0.60	0.29*	0.62	0.32
TiO_2	1.80	0.80	1.62	2.04	0.77	1.57
P_2O_5	0.95	0.46	0.07	1.39	0.18	1.04
CO_2	0.28	0.80	0.61	0.40	0.42	0.16
FeS_2	0.34	0.51	0.79	0.28	—	—
MnO	0.08	0.02*	0.04	n.d.	—	0.17
	100.29	99.56	99.93	98.73	100.02	100.32

	MODES					NORM
	A	B	C	D	E	F
Quartz	0.70	2.34	2.12	9.93	3.0	6.03
Orthoclase	5.28	1.11	—	17.53	2.0	26.21
Aibite	35.90	35.95	26.20	34.32	73.0	34.06
Anorthite	18.77	27.61	19.32	8.00		10.98
Hypersthene	4.61	2.24	—	—	7.0	11.67
Diopside	2.36	0.96	12.14	1.04		2.12
Actinolite	2.40	—	1.83	3.10	—	—
Hornblende	5.12	1.72	3.10	1.41	4.0	—
Biotite	5.48	5.19	7.95	6.74	1.5	—
Sericite	2.98	3.63	0 40	0.81	—	—
Sphene	0.78	—	0.22	0.98	tr.	—
Magnetite	6.11	1.97	5.45	5.18	2.0	2.86
Ilmenite	2.81	1.52	2.88	3.11		2.98
Pyrite	0.34	0.51	0.79	0.28	tr.	—
Apatite	2.22	1.06	0.16	3.22	0.5	2.45
Carbonate	0.58	1.68	1.04	0.90	—	0.40
Scapolite	1.15	2.94	7.25	—	—	—
Epidote	0.28	—	0.98	—	—	—
Chlorite	1.33	1.66	1.34	1.35	—	—
Kaolin, etc.	1.14	8.78	7.90	1.29		—
Garnet					7.0	
	100.34	100.87	101.07	99.19	100.0	99.76
Plagioclase	$Ab_{67}An_{33}$	$Ab_{58}An_{42}$	$Ab_{59}An_{41}$	$Ab_{82}An_{18}$	$\pm Ab_{48}An_{52}$	$Ab_{78}An_{22}$
Mafics	31.34	15.77	35.70	23.19	14.5	19.63

* Determined by J. M. Harrison.

A—Analysis of composite of 30 specimens taken at fairly regular intervals across the Tichborne intrusive. Analyst, J. M. Harrison.

Eastward—i.e., upward—it rapidly becomes coarser-grained and more anorthositic. Plagioclase ranges from An_{40} to An_{45}, and some biotite is present. Farther east the rock becomes gradually finer-grained, orthoclase occurs in notable amounts, plagioclase is less calcic, apatite and titaniferous magnetite are more abundant, sphene is locally prominent, and a little very fine-grained quartz is locally present. Thus, there is apparently a chilled lower margin, a coarse central zone, and a more strongly

TABLE 3.—*FeO : MgO ratio in dark silicates of Table 2*

	Molecular per cent	Weight per cent
A	1:2.2	1:1.2
B	1:1.2	1:0.7
C	1:2.3	1:1.3
D	1:1.8	1:1
E	1:1.75	1:1
F	1:0.73	1:0.4

differentiated upper zone in which the accessories—apatite, ilmenite, and magnetite—occur more abundantly, accompanied by more sodic plagioclase, some orthoclase, and minor amounts of quartz.

No identical conditions occur in the Tichborne mass. The dip of the foliation is not constant, but, neglecting the zones of syenite, there is a distinct variation in composition from west to east, similar to that in the Blessington. Whether the zone of syenite at the east end of the Tichborne be considered the product of differentiation in place or the result of granitization, the minerals there are characteristic of the upper layers of the Blessington gabbro. As orthoclase is present in the upper parts of the Blessington where there are no signs of granitization, some of the orthoclase in the east end of the Tichborne is probably also the product of differentiation. However, there is relatively more orthoclase and quartz in the Tichborne than in the Blessington, and the plagioclase is also more sodic. Hence it appears that differentiation of the Tichborne has produced a dioritic rock with large amounts of apatite, magnetite, ilmenite, and sphene, and lesser amounts of orthoclase. This assemblage has been modified by alaskite to produce a perthitic syenite with as much as 30 per cent orthoclase.

The Victoria gabbro appears to dip east, but relations between rock types in it are not so well known as those in the Tichborne and Blessington masses. However, except for the syenite zone along the bottom of it, the arrangement seems generally

B—Analysis of gabbroic anorthosite from central portion, west end of Tichborne intrusive. Analyst, Milton Hersey Company.

C—Analysis of gabbroic anorthosite about 200 feet nearer border of intrusive than B. Analyst, W. F. Green, Ontario Department of Mines. Analysis supplied through the courtesy of Dr. W. D. Harding, Ontario Department of Mines.

D—Analysis of pink, syenite phase at east end of Tichborne intrusive; Analyst, Milton Hersey Company.

E—Composite grab sample of Whiteface (gabbroic) facies of anorthosite, Willsboro Quadrangle, N. Y., from Buddington (1939, p. 30, Table 4 analysis 19). Analyst, R. B. Ellestad.

F—Augite-hypersthene syenite, Lowville Quadrangle, N. Y., from Buddington (1939, p. 74, Table 21 analysis 86). Analyst, T. Kameda.

similar to that of the Blessington body. The occurrence of syenite along the bottom strongly suggests that the syenite was formed by granitization, probably by unexposed alaskite.

DIKES ASSOCIATED WITH THE BASIC INTRUSIVES

Dikes are rather numerous in the vicinity of the basic masses, and their association with the basic rocks, together with certain peculiarities of composition, indicate a close genetic connection with them.

Thin sections of a few dikes near the Wilkinson anorthosite were studied. One dike, which cuts granitelike gneisses about a mile north of the anorthosite, consists of andesine (An_{47}) with about 5 per cent of accessory minerals including biotite, amphibole, pyroxene, titaniferous magnetite, and apatite. It is unquestionably anorthosite.

Some dikes in gneissic granite near the east end of the anorthosite consist mainly of oligoclase (An_{25-30}) with different amounts of biotite, titaniferous magnetite, apatite, and pyroxene. One contains fine-grained quartz along the margins, but otherwise it closely resembles anorthosite. In appearance and mineral composition these dikes are practically identical with those associated with the more gabbroic rocks around Tichborne.

Dikes of considerable variety occur in and around the gabbros. Some are lithologically similar to certain phases of the gabbro, and others seem genetically related to the gabbros. The most common dikes consist of fine- to medium-grained gabbro or diorite that weathers in various shades of brown. All are gneissic, but the lighter-colored ones are more strongly so. The brown dikes have been divided into two groups on the basis of color and degree of foliation.

The light-colored dikes, which do not cut the gabbro but are concentrated near it, are equigranular and gneissic. Minerals present are plagioclase (An_{25-35}), poikilitic amphibole, biotite, pyroxene in some, potash feldspar, titaniferous magnetite, apatite, and some quartz. All silicates are mildly altered. Composition is variable but within the limits of variation in the gabbros. These dikes are found in the Hinchinbrooke granite and in some places are somewhat deformed. They are identical with some of the finer-grained portions of the gabbro.

The darker-brown dikes are more massive than the lighter ones, and some of them cut the gabbros. They are equigranular with rounded grains of pyroxene set in a groundmass of plagioclase (An_{48-56}), which constitutes about half to three fourths of the rock. Other minerals present are amphibole, biotite, magnetite, apatite, and sphene. Microscopic examination showed two varieties of pyroxene—augite and hypersthene. Magnetite comprises about 5 per cent of the dikes, and other accessories are apatite and sphene. No potash feldspar or quartz was seen. Except for the more massive character of the dikes, greater freshness of the minerals, and more abundant mafic minerals, these dikes are so similar to the gabbros that a genetic relationship seems highly probable.

A number of dikes of other types were examined. Some of them resemble anorthosite as they consist mainly of andesine with 15–25 per cent of other minerals. One dike off the northeast end of the Blessington mass consists of andesine (An_{42}) with

15–20 per cent quartz. A few rather silicic, coarse-grained dikes occur in the gabbros and adjacent rocks. Those examined consist mainly of large grains of oligoclase (An_{15-25}) and a considerable amount of potash feldspar which occurs both as individual grains and in perthite or antiperthite. Quartz seldom exceeds 10 per cent; biotite is the dominant dark silicate but usually does not exceed 5 per cent; 5–10 per cent titaniferous magnetite is characteristic, with accessory apatite, sphene, and pyrite.

TABLE 4.—*Optical character of apatites*

	N_o	N_e	N_o-N_e	F-C(N_o)	F-C(N_e)
1.....	1.641	1.637	.004	.009	.008
2.....	1.642	1.638	.004	.009	.008
3.....	1.642	1.638	.004	.008	.007
4.....	1.638	1.634	.004	.006	.005
5.....	1.649	1.641	.008	.012	.009

1—Yellow-green apatite from central part, Tichborne gabbro.
2—Yellow-green apatite from central part, Blessington gabbro.
3—Green apatite from pegmatites in the Blessington gabbro.
4—Dark green apatite from syenite near southeast corner of Hinchinbrooke Township. Associated with some magnetite, titanite and amphibole. Sample supplied by Dr. W. D. Harding.
5—Light green apatite from McLaren Phosphate Mine, near Burridge Post Office, Bedford Township, about 6 miles east of Tichborne. Associated minerals are pink calcite, pargasite and pyrite. "Apatite occurs in veins and lenses which intrude gneisses." (W. D. Harding, personal communication.) Sample supplied by Dr. Harding.

Dikes of two rather unusual types were found only in the gabbros or in rocks immediately adjacent to them. One type, found in the Tichborne gabbro, is a black, quite massive rock consisting mainly of pyroxene and titaniferous magnetite with amphibole, some serpentine, and accessory apatite and pyrite. The pyroxenes are of two varieties—almost colorless augite and lesser amounts of pleochroic hypersthene. The pyroxenes occur as rounded grains separated from the iron ore by narrow rims of amphibole. Serpentine occurs around small grains of pyrite. In polished section the metallics are seen to be about equal amounts of ilmenite and magnetite with minor amounts of hematite and pyrite. Blades of ilmenite appear as fine lines in some of the magnetite.

The second type, consisting mainly of titaniferous magnetite, and apatite in varying proportions, cut the Blessington gabbro and the older rocks surrounding it. Some of these sinuous dikes, just north of the east arm of Eagle Lake, have been worked for apatite (Rept. Roy. Comm. 1890, p. 167; Spence, 1920, p. 44).

Cousineau (1940) described them as mainly titaniferous magnetite and apatite with tourmaline (var. schorlite), spinel (var. ceylonite), scapolite, and lesser amounts of amphibole, pyroxene, sphene, calcite, quartz, biotite, and a number of other minerals. He concluded that these dikes are pegmatitic derivatives of the gabbro.

Polished specimens of this dike contain the metallic minerals magnetite, lesser amounts of ilmenite, and minor amounts of hematite, pyrite, and two minerals that were not positively identified. Magnetite and ilmenite commonly show exsolution patterns, and where discrete grains of the two occur together the boundaries are

smoothly rounded. Hematite occurs as blades in ilmenite and less commonly in magnetite.

Optical properties of grains of apatite from these pegmatites were determined by the double-variation method and compared with those of apatite from the gabbros and of apatite from two deposits in the same general area (Table 4).

The first three specimens have optical characters so nearly alike that there can be little doubt of their chemical similarity. Presumably, therefore, they have a common origin. Specimen 4, from a pit about 2 miles west of the Island Lakes gabbro, is somewhat different. Its mineral association, however, is much the same as apatite from the Blessington mine. Specimen 5 is completely different both in mineral association and in optical properties.

Similarity of the properties of apatite in the gabbros and in the dikes supports the genetic relationship of the two. If the dikes are pegmatites from the gabbro magma, liquids rich in iron, titanium, phosphorus, and some chlorine were exceptionally abundant in the closing stages of the crystallization of the gabbros.

RELATION OF THE BASIC MASSES TO OLDER ROCKS

GENERAL STATEMENT

Detailed evidence is presented in this section to show that the gabbros intrude the Hinchinbrooke granite and have produced contact effects on it. Further, anorthosite has reacted with limestone and other rocks and has been modified by these reactions.

STRUCTURAL RELATIONS

In most places the anorthositic masses are concordant with structures in the enclosing, older rocks, but in a few places cross-cutting relationships are clear, especially where the gabbroic anorthosites are in contact with paragneisses or limestone.

Close examination is required to locate the exact contacts between gabbroic rocks and the Hinchinbrooke granite, for both the granite and the border phase of the gabbros weather to very similar, buff to brown colors.

At the east contact of the Blessington mass no sharp contact with the gneissic granite can be observed. About 500 feet from the basic mass layers of fine-grained gabbro, approximately parallel to the foliation in the granite, appear and increase in number as the gabbro is approached. Two places were found where layers of gabbro transgress the foliation of the granite at angles of 10° and 40°. About 500 feet west of the first dike of gabbro is the main mass of basic rock. The foliation both in granite and in gabbro strikes about north-south and the dip is 40°–60° E. Thus the Blessington mass is evidently an eastward-dipping sill, and the dikes of gabbro in the granite are on its upper side.

At one place on the east side of the Victoria gabbro two small dikes cut gneissic granite. These dikes are identical with typical, fine-grained gabbro that occurs about 40 feet away.

On the west side of the Victoria mass the basic rock is somewhat coarser and is bordered by typical, though somewhat pinkish, gneissic Hinchinbrooke granite that contains many inclusions oriented parallel to the foliation in the granite. At one

face amphibole gneiss, similar to material common as inclusions in the granite, is exposed on a hill side and dips about 45° NE. Above it, and in sharp contact with it, is a rather pink, anorthositic phase of the gabbro. The contact between the two rocks dips about 35° NE. (Fig. 6). Ten feet northwest from this spot the foliation in the

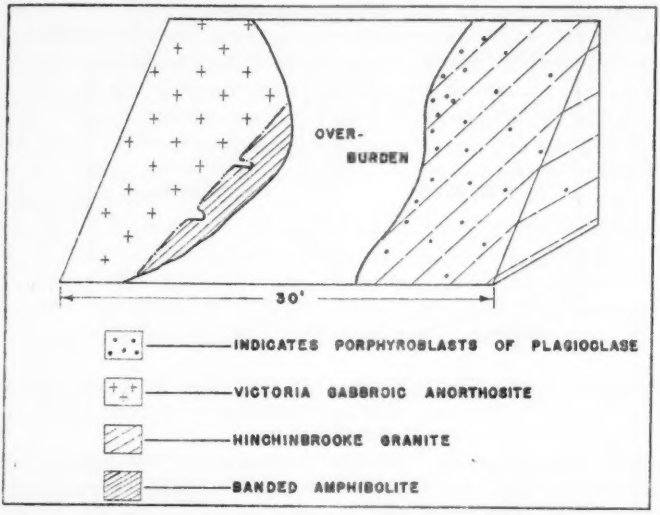

FIGURE 6.—*Sketch of west contact of Victoria gabbroic anorthosite with Hinchinbrooke granite*

gneissic granite dips 45° NE.—that is, parallel to the foliation of the amphibole gneiss. The granite contains porphyroblasts of dark plagioclase, similar to the purplish plagioclase of the anorthosite, and these are more abundant near the basic rock.

No cross-cutting relationships were found between the Wilkinson anorthosite and older rocks.

METAMORPHIC EFFECTS OF THE BASIC INTRUSIVES

In a number of places the anorthositic masses have obviously produced contact effects on the older rocks, though at most places the effects are not marked.

Limestone in contact with gabbroic anorthosite is commonly altered to a dense, fine-grained amphibolite; at the west end of the Tichborne mass, scapolite has also been developed. Limestone at the north end of the Blessington intrusive is altered largely to a mass of light, granular pyroxene. A shell of amphibole-rich rock separates the Wilkinson anorthosite from adjacent limestone.

Contact effects of gabbroic anorthosite on Hinchinbrooke granite are rather indefinite. The granite in contact with what is considered to be the upper side of the Victoria gabbro is finer-grained, less gneissic, and less silicic than usual. Toward the gabbro, the quantity of magnetite increases, becomes titaniferous, and occurs in ir-

regular, ragged grains that cut other minerals. As the gabbros were unusually rich in late titaniferous magnetite, the gabbroic magma may have been the source of the titaniferous magnetite in the gneissic granite.

The Grenville gneisses in the area under discussion are commonly feldspathic rocks containing amphibole and biotite with more or less magnetite and local pyroxene. Hence contact effects on them are not very apparent.

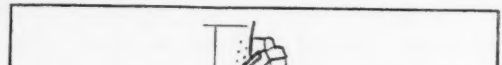

FIGURE 7.—*Sketch of inclusion in Wilkinson anorthosite*

Inclusions of various rock types occur in the anorthositic masses. Many, however, have been so altered that their original character is uncertain, indicating ability of the anorthositic magma to assimilate other rock.

Limestone inclusions in the Wilkinson anorthosite are surrounded by a shell of dense amphibolitic rock that varies in width from a few inches to many feet. This shell grades outward to typical anorthosite through a zone of gabbro by decrease in dark minerals and increase in plagioclase. The amphibolitic zone is characterized by plagioclase somewhat less calcic than that of the surrounding anorthosite. Titaniferous magnetite is unusually abundant in the vicinity of these inclusions. Amphibole is the dominant dark silicate and is usually intergrown with a mixture of chlorite and carbonate. Some biotite is present as greenish brown, rather pulpy masses. Sphene is unusually abundant, as would be expected, and apatite is invariably an accessory. In some places a little scapolite has formed.

Elsewhere in the anorthosite, rocks consisting almost entirely of amphibole and biotite with minor amounts of feldspar grade outward to typical anorthosite just as do the alteration zones around limestone inclusions. These dark rocks are probably completely transformed blocks of limestone.

Inclusions of medium-grained, dioritic or syenitic rock occur in the anorthosite. One such, on the south shore of Fish Lake, is enclosed in anorthosite consisting of coarse crystals of plagioclase (An_{47}) with only a very little magnetite, biotite, and amphibole. Crystals of plagioclase project from the anorthosite into the inclusions (Fig. 7). The plagioclase of the inclusion is oligoclase (An_{18}). Except for a little pulpy biotite the mafic minerals are completely altered to intergrowths of chlorite, carbonate, and iron oxide. The feldspars contain abundant cloudy alteration products.

Inclusions of more basic rocks are quite abundant in the gabbroic anorthosites. All, except some of those in the fine-grained border phases, are finger-grained than

the enclosing rock. Generally they grade into the host rock, which, in some instances, is true anorthosite. In other places the inclusions are transected by stringers of anorthosite that are essentially concordant with the foliation of the inclusions. Just north of Thirteen Island Lake, a very light-gray, coarse-grained anorthosite phase of the gabbro contains small inclusions of amphibolite, which are traversed by ramifying veinlets of rather fine-grained, white anorthosite. At many other places in the gabbroic anorthosites inclusions of various sizes contain greater proportions of mafic minerals than do the surrounding rocks. The mafic mineral content of the gabbroic anorthosite generally increases gradually toward the inclusion.

In a few places inclusions contain relatively little mafic mineral, and some of them contain quartz. A very few of them appear little altered and resemble the Hinchinbrooke granite, but in most cases alteration is so extreme that no original characteristics are recognizable.

Plagioclase in the more mafic inclusions is usually slightly less calcic than the host, pyroxenes are more completely uralitized, and amphibole is usually the dominant mafic mineral. The texture of the inclusions is much the same as that of the host; one inclusion has a diabasic texture. Mineralogy and texture, therefore, indicate thorough impregnation and mineral rearrangement by the gabbros.

ORIGIN OF ANORTHOSITES AND RELATED ROCKS OF THE HINCHINBROOKE-BEDFORD AREA

The Wilkinson anorthosite appears to be what Buddington called the Adirondack type of anorthosite (1939, p. 208) and described as "—masses of small to very large areal extent without obviously strongly differentiated characters" in contrast to facies of differentiated stratiform sheets or lopoliths.

The Wilkinson anorthosite is bounded in large part by syenitic or dioritic rocks petrologically very similar to the syenites and diorites associated with the anorthosites of the Adirondacks. In one or two places the anorthosite at Wilkinson is in contact with limestone.

Zones containing abundant mafic minerals, included in anorthosite, have been interpreted by Balk as gabbros "in statu nascendi" (1930, p. 297). Buddington (1939, p. 44–46) examined some of these localities considered by Balk to be basic segregations and concluded that some of the more basic zones represent thoroughly reorganized and transformed inclusions of limestone, though he admitted that local gabbro facies occur as segregations in anorthosite. The Wilkinson anorthosite contains inclusions of crystalline limestone surrounded by zones of mafic material similar to zones in the Adirondacks ascribed by Buddington to the alteration of limestone and by Balk to basic segregation.

The gradation from anorthosite to syenite at Wilkinson suggests contemporaneity. The anorthosite, however, shows a finer-grained border toward the syenite, indicating either that the syenite is older or that it was intruded along or near the original edge of the anorthosite. Phenocrysts of plagioclase in the syenite vary considerably in composition; they may have been included in the syenite magma for different periods, some long enough to be made over to plagioclase like that of the host, and others only long enough for partial equilibrium to be attained. If that be true, xenocrysts were

torn away from the anorthosite at various stages during intrusion of the syenite magma, and the syenite is therefore later. Also, blocks of anorthosite surrounded by syenitic rocks at the southwest end of the main anorthosite (Fig. 3) may have been engulfed by the syenite magma. Just south of Fish Lake some streaks of syenite in anorthosite may be intrusive, though no sharp contacts were observed. On the whole, the evidence indicates that the syenite was a little later than the anorthosite.

In some places the anorthosite is cut by small, fairly regular dikelets and stringers 1–2 inches wide, many of which appear to fill joint planes. They are sugary and in general exhibit the characteristics of aplites or pegmatites. (*See* section on Petrography of Anorthosites.) Hence, they are probably direct offshoots from the anorthosite (Balk, 1937, p. 41).

All the dikes apparently related to the Wilkinson anorthosite, and even some granite dikes that cut anorthosite, show local granulation. Thus some, at least, of the granulation of the anorthosite is due to post-consolidation movement.

In summation, an anorthositic liquid is indicated here by the presence of stringers of anorthosite in inclusions, the high degree of alteration of limestone inclusions, and the strewing of the alteration products of these inclusions through the mass, the occurrence of a dike of typical anorthosite a mile from the anorthosite mass, and the fact that crystals from the anorthosite project into inclusions. As the marginal rocks are more basic, the original magma must have been more basic than the average of the main mass of the anorthosite, perhaps with a composition about that of gabbroic anorthosite (Table 2, E) as suggested by Buddington (1939, p. 216).

One of the main objections to the possibility of a magma of anorthositic composition is the high temperature necessary to keep it molten (Bowen, 1928, p. 189). Certainly volatiles such as water, phosphorus compounds, chlorides, etc., would lower the freezing point, and probably iron and titanium oxides and alkalis would lower it still further. A temperature of 1200°C does not seem unreasonable. Crystallization of a granite magma which contained only 1 per cent water would begin at about 1025°C at a depth of 10 kilometers, but at 700°C there would still be about 15 per cent residual liquid (Goranson, 1932, p. 234–235). If a comparable condition exists in magma of gabbroic anorthosite, crystallization could continue to a temperature of about 900°C.

At Wilkinson there is evidence of volatiles in the presence of late dikes carrying quartz, tourmaline, and biotite, the fact that the pyroxenes are partly changed to amphiboles, the occurrence of late magnetite and apatite, and the alteration of limestone to assemblages of amphibole, biotite, and chlorite. However, if the magma was originally gabbroic anorthosite the mafics must have been largely removed. No segregations of basic material were seen at Wilkinson except near the margins where the anorthosite may have been contaminated by assimilation of limey rocks so that most of the mafic minerals must either have settled below the zone of observation or have been carried upward by volatiles and removed by erosion.

The masses of more gabbroic composition that occur around Tichborne, Eagle Lake, etc., present some special problems, which, however, are directly connected with the anorthosite problem as a whole. The gabbros are considered to be related to the anorthosite at Wilkinson for they contain the same minerals as the anorthosite and have anorthositic facies. Pegmatite dikes identical in mineralogy are associated

with both the gabbros and the anorthosite, and at least one dike of anorthosite is associated with the gabbroic masses.

Any hypothesis for the origin of the gabbroic anorthosites must take into account the following facts:

(1) Foliation is marked in the border zones of the gabbros but not in the coarser-grained central portions.

(2) Some granulation in the minerals of the gabbros has occurred.

(3) In general there is a gradation from anorthosite to gabbro to diorite.

(4) An unusually large amount of apatite, magnetite, and ilmenite occurs in the masses as a whole, and a large proportion of these minerals is concentrated in the upper portions of the masses.

(5) Inclusions of Grenville sedimentary rocks occur in all stages of assimilation in the gabbros, from well-defined rocks to thoroughly disintegrated masses consisting only of patches with abnormal amounts of mafic minerals.

(6) Pegmatite dikes with a very large proportion of iron oxides, apatite, and mafic silicates are related to the gabbros.

(7) Granodiorite pegmatites and dikes of hypersthene gabbro (norite) also appear to be genetically related to the gabbros.

The variation in composition of the plagioclase in the Tichborne gabbro is related to the position of the rock in the intrusive, at least in the western part of it and the northern part of the east half (Fig. 5). The finer-grained margins contain more sodic plagioclase than does the interior of the mass, and presumably represents a "chilled" zone.

The eastern half of the Tichborne intrusive shows greater irregularities in composition than the western half. However, the high percentage of potash and quartz is probably due to action by alaskite though some orthoclase appears to have been formed by differentiation in the gabbros themselves. It is not possible to distinguish between the two types quantitatively.

There is considerable evidence that volatiles were abundant in these basic rocks—more than 2 per cent of apatite for the Tichborne intrusive as a whole and more than 3 per cent in the more silicic and sodic end, at least .4 per cent combined water in the rock, the presence of scapolite, a mineral carrying chlorine, the evidence of alteration caused by the gabbros in the adjacent rock, the presence of magnetite-apatite dikes associated with the gabbros, and the fact that inclusions are thoroughly reconstituted. Moreover, the presence of nearly 9 per cent of magnetite and ilmenite, which formed after most of the silicates, indicates the presence of volatiles in the end stages of intrusion.

Broderick (1935) appealed to transfer by gases to explain the concentration of ilmenite, magnetite, and apatite in the upper parts of differentiated lava flows in Michigan. Similar processes may have been active during the crystallization of these anorthositic magmas for the anorthositic portions of these masses appear to occur near the bottom of the various gabbros. It is possible too, that some of the mafic silicates, which are present in larger amounts in the upper portions of the gabbros than they are in the lower, were formed by reaction of the iron-rich residual portions of the magma with inclusions of limey sediments.

The earliest dikes associated with the gabbros are quite strongly foliated, whereas the latest members, which cut the gabbros, are almost completely massive, suggesting that some of the granulation of the minerals in the gabbros is due to tectonic movements which lessened in intensity toward the close of the intrusive period. This would also explain the foliated borders of the masses.

SUMMARY

The Wilkinson anorthosite and the various gabbros are intrusive, derived from a gabbroic anorthosite magma containing considerable quantities of "mineralizers" and of ferromagnesian and potash constituents. Thus the temperature of crystallization was not unreasonably high. Volatiles rose in the magma chamber, apparently carrying much iron and titanium with them. As the volatiles rose the magma of the lower part of the chamber crystallized with labradorite as the dominant constituent, and, concurrently, iron, titanium, phosphorus, chlorine, etc., became increasingly abundant in the upper part. Inclusions caught in the magma were attacked, and, as the more volatile portion of the magma contained the most iron, ferromagnesian silicates were formed by reaction with lime and magnesia in the limestone and limey sediments. Assimilation of the adjacent rock and of inclusions continued until the magma had completely crystallized and the volatiles had "boiled off." On the basis of this hypothesis, assuming the gabbros and the Wilkinson anorthosite are products of the same magma, the gabbros represent a higher level in the earth's crust and a more completely differentiated product than the anorthosite. Their composition is probably considerably affected by reason of assimilation of limestone and limey sediments. This would also result in a patchy distribution of rock facies.

The source of this magma, with a composition of gabbroic anorthosite, is a matter of speculation. Crystallization from a gabbroid magma seems untenable. Selective melting of a layer in the earth's crust that consisted of bytownite anorthosite (Buddington, 1939, 1943) appears to offer fewer difficulties than any other explanation so far advanced.

The relations of the gabbros to the Hinchinbrooke granite indicate that they are younger than the granite. The alaskite is unquestionably younger than the gabbros. Hence, two batholithic granites, separated by the intrusion of the gabbros, can be distinguished in the Hinchinbrooke-Bedford area.

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EXPLANATION OF PLATE 1

PHOTOMICROGRAPHS OF CERTAIN ROCK TEXTURES

Figure

1. Photomicrograph of massive alaskite. (Crossed nicols. ×20.)
2. Optically parallel blebs of quartz (white) in microcline. The quartz blebs have the same orientation in adjacent grains of plagioclase. Alaskite. (Crossed nicols. ×27.)
3. Optically parallel, but discontinuous, grains of quartz replace plagioclase of Wilkinson anorthosite. (Crossed nicols. ×16.)
4. Stringer of stained orthoclase (dark gray) cuts across plagioclase of Wilkinson anorthosite. ×17.
5. Stained antiperthite. Orthoclase (dark gray) cuts plagioclase (light gray). Tichborne gabbro. ×83.
6. Irregular dendritic, but optically continuous, scapolite (white) transects diversely oriented grains of plagioclase. Tichborne gabbro. (Crossed nicols. ×17.)

FIGURE 1

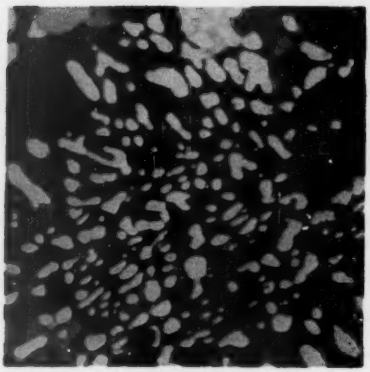

FIGURE 2

FIGURE 3

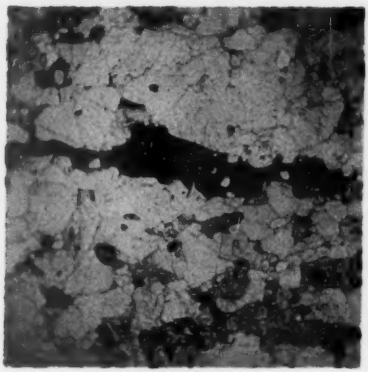

FIGURE 4

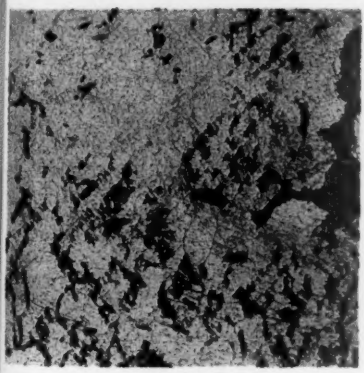

FIGURE 5

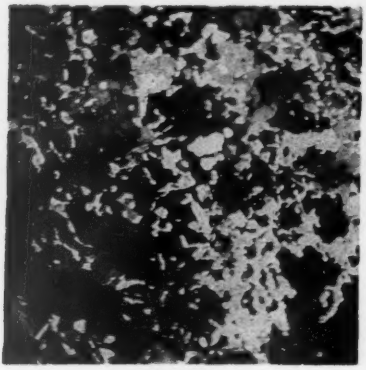

FIGURE 6

PHOTOMICROGRAPHS OF CERTAIN ROCK TEXTURES

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CLASTIC DIKES OF THE COLUMBIA BASIN REGION, WASHINGTON AND IDAHO

BY R. L. LUPHER

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ABSTRACT

Pleistocene lake and stream deposits of southeast Washington and the adjacent portion of Idaho are cut by innumerable clastic dikes of an unusual kind. Dikes are most abundant in the Touchet beds and scabland deposits, a few are present in the earlier Clarkston deposits, and rarely, they enter the Columbia River basalt. The processes of formation were long-continued and concurrent with proglacial deposition. An analysis of the sedimentary environment indicates that ice played an important rôle in deposition of dike-bearing deposits.

Many dikes occur in compound units of several or many individuals lying parallel in the same major fissure; they were formed by repeated filling of growing fissures. Fissures were filled from above by one or more of at least four processes; most of the sediment was swept into fissures by streams, lake currents, and waves, some collapsed from fissure walls or poured in from unconsolidated surficial deposits, a small amount was carried by underground currents, and films of clay and silt were spread upon fissure walls in several ways. Wind-borne sediments are indicated but not proven. Two or more processes generally operated together or alternately to fill each fissure; dike material, ranging from clay to fine gravel, may be sorted or unsorted, stratified or unstratified, and the stratification may be horizontal, inclined, or vertical. Any or all possible combinations of the variables may be present even in individual dikes.

Five processes of fissure development are recognized: (1) uneven settling and cracking through melting of buried ice, (2) gravity sliding and faulting on inclined zones of subsurface melting, (3) formation of cavities where ice blocks and layers melted, (4) erosion by underground streams, and (5) faulting and fissuring by landslides in the Columbia River basalts.

INTRODUCTION

Proglacial stream and lake deposits of the Columbia basin region are traversed by innumerable fissures which contain sedimentary fillings or "clastic dikes." The dikes are very unlike those described by Diller (1890) and Newsom (1903) for they are not characterized by a prevailing type of structure and origin or by a restricted range of composition. Instead they are complex in composition, in structure, and in mode of formation; even the fissures were of more than one origin.

The present study was made in order to determine the genesis of the dikes and of the fissures. This cannot be done from a study of the dikes alone, for they are related to a peculiar set of proglacial conditions. Therefore considerable attention will be given to the nature of the Pleistocene sediments and related physiographic processes, as well as to the dikes and fissures.

Published data on these dikes are scant. Jenkins (1925a), in a general paper on clastic dikes of eastern Washington, described dikes near Lewiston, Idaho, and near Touchet in the Walla Walla valley of Washington (Fig. 1). He concluded that cracks of earthquake origin were filled from above and suggested that the dikes originated by injection or compression when cracks opened beneath the mud of a Pleistocene lake bottom; experiments were made (1925b) with plastic clay to show that parallel layers could be squeezed into a crack between wood blocks. Flint (1938) mentioned the dikes of the Touchet and Pasco regions and furnished much information on the deposits in which dikes are present. Like Jenkins he suggested

a filling process of "flowage by squeezing" but recognized the possibility of repeated opening of the fissures to permit addition of new material. Allison (1941, p. 62) refers to "injection of silt dikes" in terrace materials along the lower Snake River. Much has been written about the deposits in which the dikes occur, especially by Bretz, Allison, and Flint, and this material contributes much to an understanding of the sedimentary environment in which the dikes were formed.

The writer was assisted throughout the field work and in the preparation of the manuscript by his wife, Anna Woodward Lupher; her clear memory of detail has constantly directed attention to new phases of the problem. Arthur Ritchie and Lorin Clark also gave valuable field assistance during the spring of 1940.

SEDIMENTARY ENVIRONMENT UNDER WHICH FISSURES AND DIKES FORMED

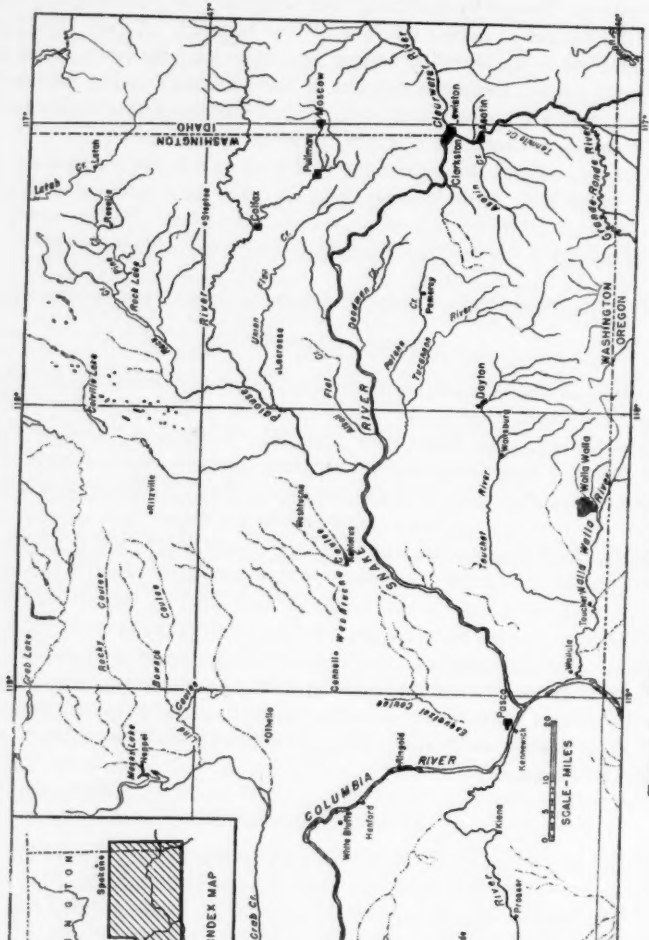
GENERAL STATEMENT

Clastic dikes of the type discussed here apparently are restricted to proglacial stream and lake deposits except in certain rare occurrences where they reach into basalt lavas beneath. No dikes have been seen in other sedimentary deposits above the Columbia River lavas even though those deposits include fluvial, lacustrine, and loessial materials ranging from the widespread Ringold formation of the early Pleistocene to restricted postglacial valley and canyon fills.

Most of the dikes were formed when streams of melt-water from the front of the continental glacier west of Spokane poured southward across the Big Bend of the Columbia, developed many rocky "scabland channels" in the Columbia River lavas, and entered ponded water in the lower basin and canyon region of southern Washington. The discontinuous deposits of gravel and coarse sand associated with the scabland tracts will be referred to as the scabland deposits; the silts and fine sands of the ponded water areas have been named the Touchet beds (Flint, 1938). A few dikes are present in Clarkston deposits (Lupher, 1944) which are older than the scabland-Touchet deposits.

CLARKSTON DEPOSITS

Stream and lake deposits older than the scabland-Touchet stage are well developed in the Lewiston-Clarkston region about 100 miles above the mouth of Snake River (Fig. 1). The deep canyons are interrupted there by the Lewiston Basin, a structural and erosional depression that reaches a depth of 2000 feet near the junction of the Snake and Clearwater rivers. Immediately southwest of Clarkston, an abandoned course of Snake River is filled with gravel to a depth of 425 feet. In the central part of the basin, adjacent to Clearwater River, gravels and sands were deposited to a depth of at least 300 feet, but only banks of hillside rubble were seen at higher levels. Tammany Creek, a small tributary of the Snake, was ponded by the Snake River fill, and its canyon was partly filled with sand, silt, and blue clay. The highest fill remnants lie at an altitude of 1160 feet above sea in the vicinity of Clarkston and show a westward inclination, perhaps largely because of the original gradients of the streams. The Clarkston gravels, except near canyon sides and in local tributary canyons, are composed of nearly equal amounts of basalt and pre-Tertiary


WASHINGTON
IDAHO
OREGON
COLUMBIA RIVER
SNAKE RIVER
INDEX MAP
SCALE - MILES
Moscow
Pullman
Lewiston
Clarkston
Dayton
Walla Walla
Pasco
Kennewick
Othello
Ephrata
Grand Coulee
Coulee City
Grand Coulee Dam
Soap Lake
Moses Lake
Ritzville
Cheney
Spokane
Seattle
Yakima
Sunnyside
Prosser
Toppenish
Selah
Wenatchee
Okanogan
Republic
Colville
Davenport
Odessa
Connell
Kahlotus
Touchet
Waitsburg
Dayton
Tucannon River
Tucannon
Palouse River
Palouse
Potlatch
Deary
Troy
Kendrick
Genesee
Juliaetta
Kamiah
Orofino
Grangeville
Cottonwood
Craigmont
Nezperce
Lapwai
Lewiston
Asotin
Grande Ronde River
Joseph
Enterprise
Wallowa
Wallowa River
Imnaha River
Snake River
Clearwater River
Lolo Creek
Weippe
Pierce
Orofino
Lewis Clark
Potlatch Creek
Union Flat Creek
Palouse River
Pine Creek
Tucannon River
Touchet River
Walla Walla River
Grande Ronde

igneous and metamorphic rocks from central Idaho and northeastern Oregon; most pebbles are well rounded and well sorted. The deposits are so well cemented by iron oxide and calcium carbonate that they stand in vertical or overhanging cliffs for many years. The oxide causes a prevailing brown color, and many basalt and granite pebbles are decomposed. Occasional large angular erratics imbedded in stream gravels, indicate that the Clarkston fill accumulated during a glacial stage.

The Clarkston stage is probably pre-Wisconsin. The degree of cementation of the fill is comparable to that of the early Pleistocene Ringold formation. Moreover, the fill persisted long enough for the canyon slopes above it to be lowered considerably by erosion and mantled with decomposed rock and soil, and to this must be added the time required for the removal of the fill from the canyons and the Lewiston Basin.

Clastic dikes have not been seen in the stream gravels of the Clarkston fill, but they are in the siltstones and sandstones, in deposits of hillside rubble, and in ice-rafted siltstone erratics incorporated in the younger scabland gravels (Pl. 1, fig. 3).

Stream gravels older than the scabland-Touchet deposits are widespread in southeastern Washington and northern Oregon, and some are probably equivalents of the Clarkston deposits. Allison (1933; 1941) states that terraces of an extensive gravel fill are widely developed along the lower Snake River, Yakima Valley, and lower Columbia River. They reach an altitude of 1150 feet on the lower Snake River and decline to 200–300 feet near Portland, Oregon. Some of the deposits were recognized as older than the Touchet beds by Bretz (1925, p. 109–110; 1928, p. 681; 1929, p. 420, 506) and Flint (1938, p. 516–619), but Allison (1941) believes that they failed to differentiate between these deposits and the younger scabland-Touchet deposits along the lower Snake River and in some of the scabland tracts.

SCABLAND-TOUCHET DEPOSITS

General statement.—The reader is referred to papers by Bretz (1923, *et seq.*), Allison (1933; 1941) and Flint (1938) for more detailed descriptions of the scabland-Touchet deposits than can be given here. These writers were concerned primarily with the origin and history of the scabland tracts, and they have produced three widely divergent interpretations. The immediate purpose here is to ascertain the conditions of sedimentation during this proglacial episode in which numerous clastic dikes were formed. This can be done to a certain extent by considering only the sedimentary features of the deposits and certain inferences on which there is general agreement, but it involves also evaluation of the three scabland interpretations because they predicate markedly different environments of sedimentation.

The scablands and their associated deposits are best developed in the coulee region of central Washington (Fig. 1) but continue eastward along the lower Snake River into Idaho and southwestward down the Columbia to the Columbia Gorge. The deposits are largely fine gravel and coarse sand; silt, fine sand, and coarse gravel are of very minor amount. They are discontinuous deposits that lie in or near scabland tracts eroded into the Columbia River basalts. The floors of main scabland channels are generally free of scabland deposits, but back in the mouths of tributary canyons, on the sides of channels, and even on interstream divides several

hundred feet above the floors, the sands and gravels are abundant. The scabland deposits are almost entirely locally derived basalt pieces except along the Snake River where they contain a large porportion of metamorphic and igneous rocks from the pre-Tertiary rocks of central Idaho. In contrast to the older Clarkston gravels, most of the scabland deposits are much less stream worn, slightly consolidated, gray or dark gray, and unweathered.

The Touchet beds are largely silt and fine sand in which are many ice-rafted erratics. With few exceptions they lie below an altitude of 1150 feet and, therefore, are most widely preserved in that great structural depression of southern Washington in which lie the Walla Walla Valley, Pasco Basin, and Yakima Valley, but they extend far into tributary valleys and canyons, especially the Snake River canyon and its tributaries. The Touchet beds have been regarded as ponded- or slack-water deposits contemporaneous with, and grading into, the coarse scabland deposits. The existence of this body of water, whatever its origin, is generally admitted, and it is commonly referred to by Symons' (1882) designation of "Lake Lewis."

Clastic dikes are abundant in both types of deposits but reach their greatest development in the Touchet beds where almost every good exposure reveals dozens or even hundreds of dikes. Therefore these deposits offer the best opportunity for estimating the peculiar set of conditions responsible for the growth of fissures and dikes.

Scabland hypotheses.—Bretz (1923, *et seq.*) regarded the scabland deposits as little-modified bars left by a great rush of water, the "Spokane flood," which he believed was also the cause of the scabland tracts. The Touchet beds were attributed to deposition in ponded flood water that rose to an altitude of about 1150 feet above sea in the basin and valley region south of the scablands because of the narrow southern outlet at Wallula Gap on the Columbia River.

The beginning of an alternative scabland explanation is evident in the work of Hodge (1931) who described some local scabland effects adjacent to moraine deposits near Arlington, Oregon, and attributed them to erosion adjacent to ice jams in a former course of the Columbia River. The ice-jam concept was applied more generally to scablands in Washington and Oregon by Allison (1933) who believed that scabland deposits and scabland erosion were products of large streams working in conjunction with ice jams which clogged proglacial channels and repeatedly diverted the streams into new courses. Likewise the Touchet beds were attributed to ponded water formed by ice jams, perhaps aided by lava flows or landslides, in the Columbia Gorge. Allison stated that much of the gravel attributed to the Spokane flood by Bretz is part of a widespread older fill. Hodge (1934) stated that the scablands have had a complex history and postulated a long list of late Tertiary and Pleistocene events involving three glacial advances, ice jams, stagnant ice, spreading of loess, and various other factors, but he did not present evidence to support his views. A preliminary statement by Hobbs (1943) has some similarity to that of Hodge, particularly in stating that the ice sheet advanced far into the scabland region, but this too awaits presentation of supporting evidence.

Flint (1938) concluded that scablands and scabland deposits were formed during

the growth and dissection of an extensive fill along the lower Snake River and in the scabland tracts marginal to a rising body of water in which the Touchet beds were deposited. He argued for the existence of normal streams of proglacial water, and his analysis of the scabland deposits of the Cheney-Palouse tract presents much sound evidence against the flood concept of Bretz. The fill hypothesis of Flint likewise is contradicted by the several lines of evidence presented by Allison (1941) who reiterates his earlier (1933) contention that there are two sets of Pleistocene valley deposits along the Snake and Columbia rivers and in scabland tracts to the north. He believes that Flint, like Bretz, failed to differentiate these two deposits and so based much of his evidence for the fill during the Touchet-scabland episode upon the remnants of a much older gravel fill that is here referred, in part, to the Clarkston episode. The distinction between the two sets of deposits is accepted in this paper, and the terms of Touchet beds and scabland deposits are employed for the fine and coarse deposits of the younger episode.

Scabland-Touchet deposits along the lower Snake River.—The critical area described by Allison (1941) along the lower Snake River near its junction with the Columbia is said to show the little-consolidated scabland deposits mostly as a discontinuous mantle of gravels and coarse sands along the sides of the canyon. The deposits lie in or near abandoned channels and scabland tracts that were eroded by streams flowing along the sides of the canyon as much as 700 feet above the present river bed. In some places the scabland deposits mantle eroded surfaces of older terrace remnants and include reworked materials from the terrace gravels, but toward the bottom of the canyon the coarse sands and gravels are generally lacking and locally grade into the silts and fine sands of the Touchet beds which form a low terrace near river level. These conditions indicate strong currents high up along the sides of the canyon and slack water in the bottom of the canyon. Allison (1941) infers that the Touchet beds in the canyon are the product of Lake Lewis and that the arm of Lake Lewis which extended up Snake River was blocked by ice

> "so that the water of the river was forced to flow along the margins, and thus, impinging against the canyon walls, the currents developed scabland, modified the terrace remnants, cut channels outside the canyon walls where spillovers occurred, cut back Palouse scarps, and otherwise produced an array of features not explainable either by Bretz's Spokane flood or by Flint's fill hypothesis."

Touchet silts and sands, the slack-water equivalents of the scabland deposits, extend for many miles up the tributary canyons of the Snake that lie east of the Palouse River.

Scabland-Touchet deposits of the Lewiston Basin region.—The scabland and Touchet deposits continue eastward along the Snake River into the Lewiston Basin region; they have been described by Bretz (1929, p. 417–427, 505–509) who attributed them to a great rush of water generated by the Spokane flood and forced eastward up the Snake River. Discontinuous deposits of gravel and coarse sand are abundant near the river and extend to an altitude of 300 feet above the river. Upon the gravels in some places but mostly at higher altitudes and far back in tributary canyons are silts and fine sands of the Touchet type which lie as an undulating mantle upon the dissected surface of the Columbia River basalts and remnants of the earlier Clarkston

fill. They reach thicknesses of 10 to 25 feet near the lower levels of the basin (Pl. 1, fig. 2), thin upward so that they cover the Clarkston terrace gravels at the 1150-foot level to a depth of only a few feet, and are seldom discernible above an altitude of 1200 feet. Ice-rafted erratics are abundant in and on the Touchet mantle and extend beyond to an altitude of 1325 feet. Most of the Touchet is crudely stratified, but some shows a peculiar rhythmic bedding that is characteristic of the Touchet in other regions (Pl. 3, fig. 1). Here, then, is a repetition of the condition described by Allison (1941) along the lower Snake River—remnants of an early stage of Pleistocene aggradation veneered with deposits of the Lake Lewis stage.

The gravels and coarse sands of the scabland type and related erosional features verify Allison's conclusions regarding ice-clogged channels. Along the Snake River canyon, near Lewiston, banks of Snake River gravel and sand (Pl. 1) lie in the mouths of most tributary canyons. They reach altitudes of more than 200 feet above the river but do not extend more than half a mile back in the canyons. These deposits contain numerous ice-rafted erratics and are gray and little-consolidated like other scabland deposits. Basalt pieces predominate in the gravel, but plutonic and metamorphic rocks from central Idaho and northeastern Oregon are present in amounts that range from 10 to 48 per cent in the stream-worn pebbles, and probably more than 80 per cent of the recognizable ice-rafted erratics are foreign rocks whose source can be traced with certainty to the pre-Tertiary exposures south and east of the Lewiston Basin.

The basalt walls of the Snake River Canyon south of Lewiston were strongly eroded during the scabland-Touchet episode to an altitude of more than 300 feet above the river. The most prominent erosional effects are notches and rock terraces cut in the main canyon side or on spurs on the upstream sides of scabland gravel banks. Significantly, all the prominent terraces and notches are on the inside of bends in the canyon. One prominent terrace, partly covered by scabland gravel, extends around the inside of a big bend of the river north of Asotin and continues northward for a total distance of 4 miles (Pl. 1, fig. 1). It is unrelated to the common rock benches that develop by differential erosion on resistant basalt flows for it traverses thick, massive lavas of the Asotin stage (Lupher and Warren, 1942) and cuts across Columbia River basalt flows at a low angle. At Silcott, 9 miles west of Lewiston, the river was forced out of its channel and now trenches a gravel deposit to a depth of more than 100 feet and for a distance of a mile, again on the inside of a big bend.

These erosional and depositional features are significant in determining which interpretation of scabland-Touchet history reflects, most accurately, the sedimentary environment under which the numerous fissures and dikes were formed. The features require strong currents of water along the sides of the canyon some 300 feet above the present river level after the canyon had been cut to its present depth. A great bore of flood water rushing eastward up the Snake River from the main scabland area, as advocated by Bretz (1929), is ruled out by the composition of the gravels and erratics which shows that the water came down the Snake River. The fill thesis (Flint, 1938), offered to account for high-level erosion and deposition near the mouth of the Snake River but not specifically applied in the Lewiston

FIGURE 1. VIEW NORTHWESTWARD ACROSS LEWISTON BASIN FROM VICINITY OF ASOTIN
a, Clarkston gravel fill in former course of Snake River; *b*, rock terrace of scabland-Touchet stage; *c*, scabland gravel and Touchet silt lying on rock terrace at mouth of Tammany Creek.

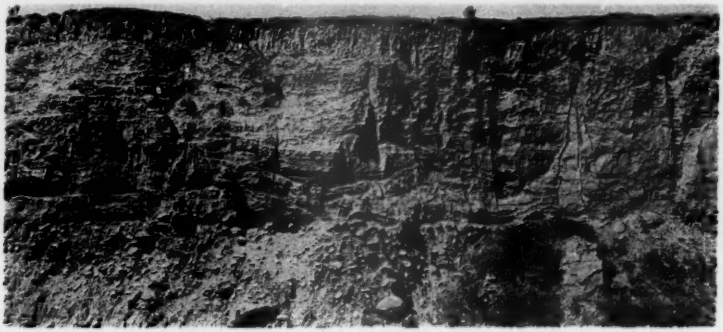


FIGURE 2. SCABLAND GRAVEL OVERLAIN BY TOUCHET SILT IN MOUTH OF TAMMANY CREEK CANYON
Dark lenses at and near base of silt may be sand-filled cavities left by melting ice; prominent lens in center of view probably a cavity that opened as silt layer below sagged on melting ice.

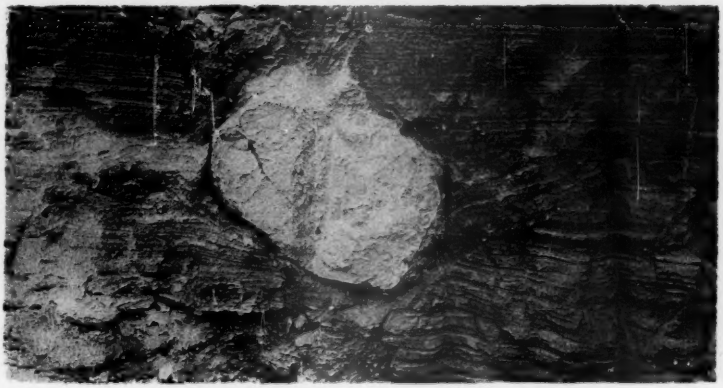


FIGURE 3. SCABLAND SAND IN MOUTH OF TENMILE CANYON
Boulder, 17 inches long, is ice-rafted siltstone traversed by compound sandstone dike of Clarkston age.

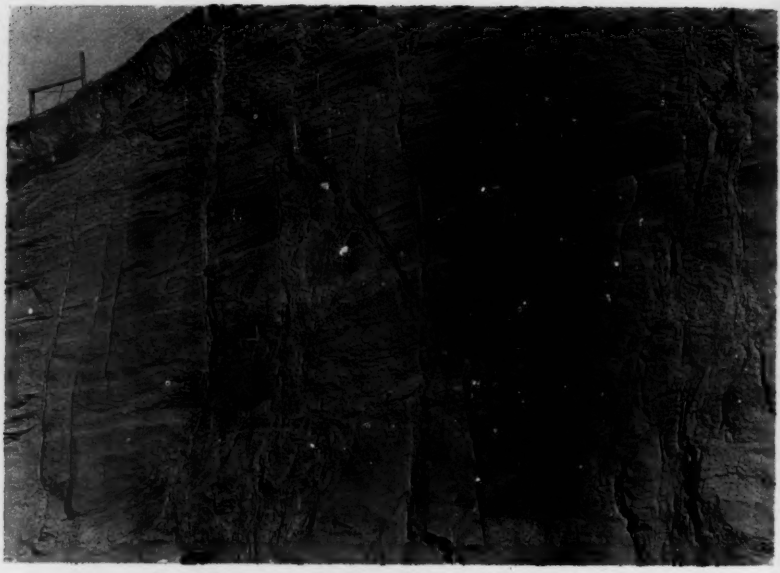


FIGURE 1. SILT AND FINE SAND 3 MILES NORTH OF TOUCHET.
Dikes are single and compound; upper part of exposure shows rhythmic bedding.

FIGURE 2. COMPOUND DIKE 2.3 MILES WEST OF TOUCHET
White scale is 6 inches long. Most dikes contain heterogeneous materials and structures. Scale lies upon 10-inch dike of unsorted sand and pebbles; on the left are about 20 small dikes; on the right is medium-sized dike with pebbles, gravel, and sand below and foreset bedded sand above.

TOUCHET BEDS AND CLASTIC DIKES

Basin, is likewise untenable because the deposits do not rise to any consistent level, and their surfaces are rounded and sloping rather than terracelike. A fill graded to the level of the upper gravels, notched spurs, and rock terraces would have occupied an area of more than 15 square miles in the middle of the basin, but not one terrace remnant remains. It seems impossible that such a surface could have been entirely destroyed by erosion while terraces of Clarkston gravel and a complete segment of canyon fill survived even the prolonged dissection that antedates the scabland deposits. Moreover, some of the scabland gravels within 60 feet of river level are covered by the mantle of Touchet silt which was spread widely over the lower portion of the basin during the scabland-Touchet episode, thus showing that the gravel deposits were little modified by later erosion.

With the Spokane flood and gravel fill concepts excluded, only the ice-jam concept of Allison (1933; 1941) remains. This explanation not only meets the necessary requirements but is also strongly indicated by location of notches and rock terraces on the inside of river bends for that would be the shortest and most logical course of an ice-blocked stream. Further confirmation is found in the work of Waters (1933) and Flint (1935) who described similar erosional and depositional features along the canyons of Columbia and Okanogan rivers in northern Washington and found good evidence that they were formed by streams flowing along the margins of ice that lay in the canyons. The continental glacier invaded that region and so offered a ready explanation of the source of the ice, but true glaciers were not present in the Lewiston Basin region.

It is not necessary to postulate local jams of glacier ice which independently raised the river level 300 feet or more, for an arm of Lake Lewis advanced, and probably periodically retreated, along the Snake River, and the only necessary function of the ice would be to clog the narrow arm of the lake so that the current would be concentrated between the ice and the canyon side or above the level of the ice. The fact that the lake water rose 175 feet higher in the Basin than it did on the lower Snake perhaps indicates the cumulative amount of clogging in the deep canyon west of Lewiston Basin. The lake water, through deep winter freezing, seems to be an adequate source of ice, but the erratics suggest that bergs from mountain glaciers were present.

Touchet beds of the southern basin region.—These deposits of silt and sand are best developed in the low portions of the Yakima, Walla Walla, and Touchet valleys; they are either absent or buried beneath younger deposits in the low central portion of Pasco Basin (Flint, 1938, p. 497) and thin or absent at higher elevations. Clastic dikes occur in greatest profusion in the Touchet beds, and these deposits may best reveal the conditions responsible for fissures and dikes.

The Touchet beds have been generally attributed to proglacial deposition in a body of water which Symons (1882) named Lake Lewis. The origin of the body of water is unknown, but three explanations have been suggested: (1) regional subsidence and sea invasion (Russell, 1893; Bretz, 1919), (2) damming of Columbia River in or near Columbia Gorge by ice, landslides, or lava flows (Russell, 1893; Hodge, 1931; Allison, 1933) and (3) impounding of flood water above Wallula Gap (Bretz, 1925). The flood concept was based largely on the scabland features and

scabland deposits, and the evidence preserved there is greatly discounted by Allison (1933; 1941) and Flint (1938). Certainly there is little evidence of a flood in the Touchet beds, and it would be difficult indeed to fit the clastic dikes to the environment of such a cataclysmic event. Choice between the other concepts is not important here, for, whatever the cause, the ponding took place and the conditions of Touchet sedimentation can be interpreted more reliably from the nature of the deposits.

The most reliable evidence for Lake Lewis is the abundant ice-rafted erratics of granitic and metamorphic rocks that are scattered over areas of lavas and young continental beds adjacent to the Snake and Columbia rivers up to an altitude of more than 1100 feet. The Touchet beds lie within the intricately patterned area of erratic distribution, and, though many of the observable erratics are upon the Touchet, some are buried in it. However, the conclusion that the Touchet is the product of ponded water apparently was influenced considerably by the presence of ice-rafted erratics, for such common lake features as fine, laminated sediments in the middle of the lake area, deltaic deposits with long foreset beds, and strandline markings are lacking in the Touchet area. This lack was noted by Russell (1893) and later workers and considered evidence for an ephemeral character of the ponding. Other abnormal features are evident in the abundance of current-bedded fine sand in flat-topped terraces resembling stream terraces.

The Touchet deposits are thickest in the bottoms of valleys and canyons and become thinner on valley slopes, especially near the upper limit of Lake Lewis, and are seldom recognizable on steep canyon sides. In some places the original surface of the deposit is flat and nearly level in the valley bottom but rises on the valley sides so that the Touchet surface roughly conforms to the pre-existent topography. Near Touchet, in the Walla Walla Valley, extensive terraces (Pl. 3, fig. 1) indicate a surface that was gently inclined with the gradient of the valley but was nearly horizontal in transverse profile for a distance of about 4 miles. The deposits are more than 125 feet thick in the middle of the valley but thin abruptly above the terrace surface so that basalt bedrock is commonly exposed on the valley sides. This condition is very noticeable along most of the tributaries of the lower Snake River where the Touchet is almost entirely restricted to erosional remnants and terraces, showing thicknesses of 20 to 60 feet, in the bottoms of the canyons and valleys; the higher slopes show mostly basalt outcrops and occasional ice-rafted erratics. Allison (1941) notes this feature along the bottom of the lower Snake River canyon in a prominent low terrace of Touchet beds, though he states that a thin marginal layer of the terrace deposits extends up the canyon sides above the terrace. Thus it appears that, while some of the Touchet deposits extend up valley slopes in a manner to be expected of deposits strewn over a lake bottom, the thickest deposits have the character of valley and canyon fills spread by streams.

The abundance of current-bedded sands in the Touchet also indicates that streams played an important part in Touchet deposition. The Touchet deposits in valley and canyon bottoms are characterized by a peculiar rhythmic bedding caused by an alternation of current-bedded and massive layers ranging from a few inches to a few feet in thickness (Pl. 3, fig. 1; Pl. 5; Pl. 6, fig. 2). The normal sequence begins

FIGURE 1. RHYTHMICALLY BEDDED TOUCHET SAND AND SILT IN WASHOUT 3 MILES SOUTHEAST OF TOUCHET
Vertical face is 130 feet high. Each layer begins with current-bedded sand and grades up to massive silt; surface of deposit forms prominent terrace. Compound dike is 120 feet deep.

FIGURE 2. RUBBLE DIKES IN TOUCHET SILT HALF A MILE NORTH OF LEWISTON
Hillside rubble has fallen into landslide fissures.

RHYTHMICALLY BEDDED TOUCHET AND RUBBLE DIKES

FIGURE 1. FORESET BEDDING IN COMPOUND DIKE 2.5 MILES NORTH OF TOUCHET

Two groups of dikes are divided by wedge of silt. Normal type of foresets in lower right; exceptionally steep foresets on left. T=Touchet silt.

FIGURE 2. DIKES AND LOW-ANGLE FAULTS 3 MILES NORTH OF TOUCHET

Dike *a* is offset above by later dikes *b* and *c*; *a*, *c*, *d*, *e* are displaced by horizontal fault; *b*, *f*, *g* are post-faulting but follow displaced dikes. Dikes on left are separated by slabs and wedges of Touchet silt.

COMPOUND DIKES IN TOUCHET BEDS

abruptly with medium and fine-grained sand which is well sorted, well stratified, and shows much small-scale cross-bedding; this is followed by a gradation to massive silt in which are many grit and coarse sand particles of basalt and also of crystalline rocks foreign to the region. A recent washout, adjacent to the Burlingame ditch near the town of Touchet, shows 38 of these rhythmic sequences in a depth of 120 feet (Pl. 3, fig. 1).

It is probable, as pointed out by Flint (1938), that some of the silt was locally derived from fine-grained sedimentary deposits that lie above the Columbia River basalt, but the Touchet as a whole contains much foreign material, ranging from large erratics to small grains of sand, which could have entered the Touchet region only along the Columbia or Snake drainage ways. Ice transportation of these materials is emphasized by their occurrence on valley slopes and in the bottoms of valleys and canyons far back from the Snake and Columbia rivers where they could not have been carried by streams. Therefore, the massive silt layers with small erratic stones probably record lake-ice deposition, and the intervening layers of current-bedded sand are largely fluvial. The common alternation of stratified sand and massive silt layers indicates a periodic advance and retreat of the lake shore, so that, in any local area, sediments were carried in as the lake level rose and reworked and sorted by overloaded streams, and probably by waves and currents of the lake margin, as the lake level fell. This process readily accounts for the great thickness of the rhythmically bedded deposits on valley and canyon floors, the graded surfaces of these deposits, and the prevalence of massive silt on the valley slopes above the reach of streams, and it accounts for the lack of strand lines, thick deltaic deposits, and other characteristic features of a long-stationary lake level. It also explains the fine grain of the silty beds here attributed to lake deposition, for an oscillating lake level would facilitate ice transportation of Snake and Columbia river silts and erratics into all the minor embayments of Lake Lewis.

Most of the Touchet shows marked small-scale deformation that does not extend to the underlying rocks. This was recognized by Flint (1938, p. 499–500) who states:

> "The Touchet beds commonly exhibit zones of warping and folding, and very commonly are cut by faults and sedimentary dikes. Intense folding, including miniature recumbent folds and overthrusts, is confined to thin zones, a few inches to a few feet in thickness, of fine-grained sediment. One of these is figured by Bretz (1928a, fig. 9). As these zones are underlain and overlain by nondeformed beds, they are the result of contemporaneous deformation, probably by slumping and sliding of water-saturated silt on gentle subaqueous slopes. Similar phenomena have been observed in glacial lake deposits in other regions. The Touchet silt and sand along the lower course of the Snake River, near its mouth, lie in broad gentle undulations affecting series of parallel beds up to 20 or 30 feet in thickness. The undulations are 20 to 100 feet long, and have amplitudes up to 10 feet. Their axes are generally transverse to the Snake River. These structures, too, may be the result of flow affecting the silt, possibly analogous to the 'mud lumps' off the mouth of the Mississippi River.
>
> Many sections, in both the silt facies and the coarser facies of the Touchet beds, are cut by normal faults with displacements up to 3 or 4 feet."

Intense folding and miniature recumbent folds and overthrusts may rightly be interpreted as structures of contemporaneous sliding, but the writer found little of this type of structure, whereas the gentle undulatory structure is present in most Touchet exposures. Many undulations are of smaller magnitude than those mentioned by Flint, and small ones are superposed on broad ones. Here and there the

gentle undulations pass into pronounced kettle-shaped sags (Pl. 5, fig. 2; Pl. 6, fig. 1). Vertical layers, recumbent folds overturned by downward movement, normal step faults, and broken and tumbled strata have been noted in these sags. The beds above normally show upward decrease in amount of downwarping, indicating that sagging continued while sediments accumulated above. Erosional channels filled by sediments swept over the sides rather than by horizontally stratified sediments (Pl. 6, fig. 1) have been noted in several places, and in cross section they resemble the downwarped sags. Some seem to be a combination of downwarping and channel fill. Undulatory bedding, sags, high- and low-angle faults, and fissures are interpreted here as largely due to the melting of ground ice. This question will be discussed in more detail.

DESCRIPTION OF CLASTIC DIKES

PRELIMINARY SURVEY

The dikes cannot be classified on the basis of physical characteristics because materials and structures are both complex and extremely variable within individual dikes. The dike material ranges from clay to gravel; it may be sorted or unsorted, stratified or unstratified; and the stratification may be horizontal, inclined, or vertical. Any or all of the many combinations of these variables may be present in a single dike. The character of enclosing deposits cannot be a basis of classification, for it bears no direct relation to the character of the dikes; gravel and sand dikes are common in silt deposits, and silt and clay dikes are common in gravel deposits. Likewise the age of a dike tells very little, for the dikes of the Scarland-Touchet episode do not differ significantly from those of the Clarkston episode. It will be necessary, therefore, to describe the full range of materials composing the dikes, the structures assumed by those materials, as well as the grosser structures or arrangement of the dikes and the fissures. The various modes of filling and origin of fissures can then be considered.

The dikes are the sedimentary filling of a maze of fissures that opened periodically during the deposition of the Pleistocene lake and stream sediments. Less than half the dikes occur as single units filling the entire fissure in which they lie; most are arranged in compound or multiple units in which two or more individual dikes lie in parallel position in the same fissure (Pl. 2, fig. 2; Pl. 4). Compound dikes of 2 to 12 individual dikes are very common, few contain more than 25, and the maximum number observed in a single fissure is 80. These dikes show periodic filling of growing fissures.

The clastic dikes, being largely restricted to little-consolidated Pleistocene sediments, are clearly revealed only in recent exposures, such as new road cuts, gravel and sand pits, and steep banks along active streams. Consequently only the widths of dikes and fissures are accurately known, the total depths of less than 25 per cent of them can be seen, and the total lengths of only a few small ones are known.

Individual dikes range in width from a millimeter or less to several feet, but dikes in the ½–6-inch range (Pl. 2, fig. 1) are by far the most common. Compound dikes are correspondingly wider, commonly between 2 and 18 inches and range up to an observed maximum of 45 inches. Probably more than half the dikes reach greater

FIGURE 1. GRAVITY FAULT 3 MILES NORTH OF TOUCHET
Rhythmically bedded silt and sand (r) faulted down about 6 feet against massive silt (m); beds above (x) are younger than fault.

FIGURE 2. SAGGING AND STEP FAULTING 5 MILES WEST OF WALLA WALLA.
Prominent compound dikes bound down-faulted block.

FAULTED TOUCHET BEDS WITH DIKES

FIGURE 1. SAG IN RHYTHMICALLY BEDDED SILT AND SAND
Strata in lower right were downwarped, faulted, and fissured coincident with deposition; inclined strata at upper left probably deposited in erosional channel.

FIGURE 2. HIGH- AND LOW-ANGLE FAULTS
Exposure is 20 feet high. Low-angle faults offset compound dike from *a* to *a'* and *b* to *b'*; dikes on right formed during or after faulting. Note vertical displacement of fissure walls.

DEFORMED TOUCHET BEDS 2 MILES EAST OF WALLULA

depth than 15 feet, and the maximum observed is 120 feet. Large compound dikes are commonly recognizable on opposite sides of highway cuts, showing lengths of 50 to 100 feet, and one compound dike in the Asotin sand pits has been traced for 300 feet.

Most dikes contrast with the enclosing deposits, for the two generally are of different grain size, and the color ranges from light buff in the clays and silts to dark gray in the coarse sands and gravels. Dikes and enclosing deposit are commonly equally resistant, though in some exposures the dikes lie in deep grooves and in others stand in high relief. The number of dikes varies with the enclosing deposit; they are most abundant in silts and fine sands, least abundant in gravel. They are especially numerous in the fine-grained Touchet beds, and almost every Touchet exposure 25 feet long shows at least one dike. It is not unusual to find 20 fissures, containing both single and compound dikes totaling 100 or more individual dikes, in an exposure 100 feet long.

DIKE MATERIALS

The dike materials range from clay to fine gravel; sand and silt predominate, but thin clay layers are present in nearly all dikes. Sorted and unsorted materials are present in a ratio of about three to one. A minor proportion of the dikes are of a uniform grain, size, and degree of sorting throughout their observable portions; most dikes that can be observed through a vertical distance of several feet show abrupt changes of grain size (Fig. 2); some sections will be unsorted mixtures, and others will be sorted or even well stratified. Here and there pebbles as wide as the dike are wedged into the fissures (Fig. 2, B).

Most of the dike materials are sediments that were swept into open fissures, and only a small amount of sediments fell from fissure walls. Therefore, marked differences between dikes and enclosing deposits are very common; dikes of gravel (Pl. 3, fig. 2) are abundant in silt beds, and dikes of fine sand, silt, and clay are abundant in gravel deposits. There is, of course, a closer correspondence of dikes and enclosing deposits where conditions of sedimentation were least variable, as, for example, in the Touchet beds where materials other than sand and silt were rarely available. Nevertheless, the dike sediments are generally coarser and better sorted than the average Touchet, for they originate largely in the current-bedded layers rather than in the massive silts. The scabland deposits, containing a wide range of materials, show many contrasts of dike and enclosing deposits; many deposits of dark-gray sands and gravels are cut by light-buff silt and sand dikes originating in overlying Touchet beds.

The dikes formed contemporaneously with the Clarkston and scabland-Touchet deposits, and so the rock and mineral composition of the dikes reflects the prevailing conditions of sedimentation. The dikes of the scabland deposits in and near the scabland channels contain basalt gravel and sand of local origin as well as fine sand and silt carried considerable distances by the melt-waters. In and near Lewiston Basin the scabland deposits contain a high proportion of sediments from the pre-Tertiary rocks of central Idaho and northeastern Oregon, and these foreign elements are also prominent in the dikes. The Touchet dikes likewise contain much foreign material characterized generally by abundant mica flakes.

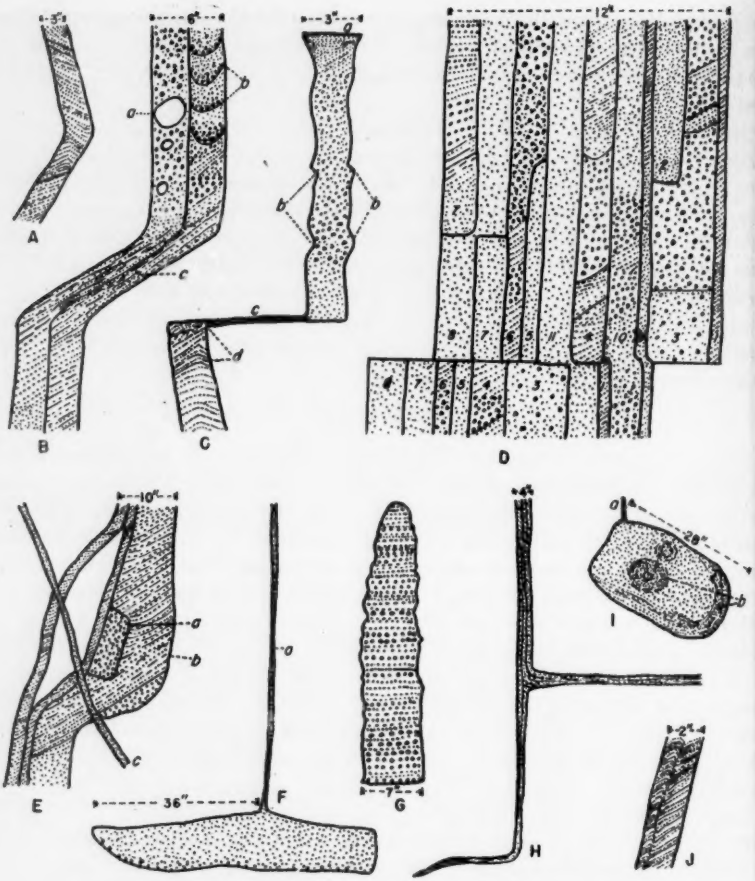


FIGURE 2.—*Sketches of clastic dike sections to illustrate structures and materials*

A. Control of dip of foresets by dip of fissure. B. Compound dike. *a* pebble, *b* troughlike stratification formed by periodic infall of sand and pebbles, *c* parallelism of stratification and fissure walls on segment of low dip. C. *a* Top of fissure, *b* fissure walls unmodified by collapse, *c* horizontal dike segment resembling fault displacement, *d* overhanging ledge and resultant stratification. D. Compound dike displaced by horizontal fault and fault cut by later dikes; possible sequence of dikes is shown by numbers. Lower part of 2 has collapsed into fissure of 3; 5 is either a remnant of collapsed dike or record of partly filled fissure. E. *a* Remnant of collapsed dike, *b* fissure widened by collapse of wall material, *c* later dike. F. Sand-filled cavity formed by melting of ice lens. *a* Feeder dike. G. Tunnel; cut and filled by subterranean stream. H. Sills and vertical separation of sedimentary layers. I. Sand filled cavity formed by melting ice block. *a* Feeder dike, *b* concentric stratification parallel to surface of melting. J. Deformation of dike sediments by movement of fissure walls.

STRUCTURAL ARRANGEMENT OF DIKES AND FISSURES

Pattern in vertical and horizontal sections.—The dikes show a great diversity of pattern in vertical section. Some are nearly straight and parallel, and either closely spaced (Pl. 2, fig. 1) or distantly spaced (Pl. 3, fig. 1); others are variously oriented,

and many are of irregular, sinuous, and zigzag pattern (Pl. 2, fig. 2). Most dikes are nearly vertical; dips of 60° are not uncommon but dips of less than 45° are rare. This applies to the orientation of the dikes as a whole, not to restricted parts of them, for sinuous and zigzag dikes of steep dip commonly have short sections of low dip; even short horizontal sections are present, causing a superficial resemblance to displacements by horizontal faults (Fig. 2; Pl. 4, fig. 2).

In horizontal section the dikes trend in all directions and show no greater abundance in one direction than another. Exposures are largely restricted to steep, fresh surfaces so that one can seldom trace their areal pattern more than a few feet. In general the plan is similar to the section, showing fairly consistent trends as a whole with minor irregularities. A large compound dike in coarse sands exposed in pits at the south edge of Asotin is somewhat irregular in detail but does not deviate more than 15 feet from a straight line for a distance of 300 feet. Likewise dike exposures in opposite highway cuts 50 to 80 feet apart commonly show little difference in strike.

Compound structure.—Perhaps the most striking feature of many dikes is a pronounced layered structure parallel to the fissure walls (Pl. 4). Much of Jenkins' (1925a) discussion of the dikes is beclouded by the erroneous interpretation of compound dikes as vertical stratification or layering of a single episode. His experiments (1925b) attempted to explain simultaneous introduction of several layers into a fissure. The true nature of these structures seems to have been recognized by Flint (1938, p. 500), for he mentions the possibility of "repeated re-opening of the fissures to permit the addition of new material." It is not difficult to prove that the parallel layers are but individual adjacent dikes standing within fissures that opened and filled periodically or opened continuously and filled periodically. Examination of compound dikes (Pl. 4) shows little to suggest the squeezing of plastic layers into an open fissure as proposed in the laboratory experiments of Jenkins (1925b). Those experiments would require that the same layers occur in reverse order in opposite halves of the dike. Instead, we find compound dikes of two, four, or more layers of different composition and different thicknesses; we find dikes of three, five, or seven layers that cannot be halved; we find dikes that come down between two earlier dikes, offset abruptly, and continue downward between another pair of earlier dikes (Fig. 2, D; Pl. 4, fig. 2). Moreover, the layers of gravel (Pl. 3, fig. 2) and even talus breccia are hardly materials that could be squeezed into fissures in unconsolidated silts.

The compound character of many dikes shows that dike formation was a long-continued process. Evidence that dikes were formed through much of or all the episode of proglacial deposition is found in the dike intersections, in which single and compound dikes cut sharply across earlier dikes, and in the dikes that originate and pass downward from various levels of the deposits.

Lateral offsets.—Most dikes are sharply defined by thin seams of buff clay and silt on the walls of fissures. This distinctness of individual dikes makes it possible to trace most of them in their sometimes intricate courses. Many dikes are offset laterally as if displaced by small horizontal faults. Some are truly offset by faults (Pl. 4, fig. 2; Pl. 6, fig. 2), but more commonly the apparent displacement is caused

by the original course of the fissure which followed first between one and then another pair of older dikes (Pl. 4, fig. 2). In this latter case the horizontal offset of the older dike, or dikes, will be equal to the width of the new fissure, and the course of the new dike will show a corresponding offset in the opposite direction (Fig. 2, D). Offsetting is not confined to units within compound dikes; many single and compound dikes are horizontally offset a few inches or a few feet along well-defined bedding planes which permitted the fissure to deviate abruptly from its otherwise regular course (Fig. 2, C). In most cases the horizontal sections are relatively thin because of the dominantly horizontal movement of the fissure walls. Units of compound dikes normally occur in greatly different order on opposite sides of these horizontal offsets. Many dikes appear to branch around lenticular slabs of the enclosing deposits or diverge at angles up to 90° (Pl. 4), but this, too, is commonly the result of a series of single fissures which, opening in succession, failed to follow exactly the same pattern as previous fissures.

Variations in width.—The compound dikes demand an unusual condition in which newly formed dikes did not collapse as a new fissure opened beside them but stood in steep or overhanging walls while sediments fell or cascaded from above. Most fissure walls are little modified by collapse or attrition. As a rule the fissures show few abrupt changes in width, and small projections on one wall are generally matched by corresponding re-entrants on the opposite wall. Many fissures, however, have been modified by removal of material from their walls. This is indicated by lack of exact correspondence of opposite walls and by more abrupt and marked changes of width (Fig. 2, E). The abrupt changes occur mostly in compound dikes where sections of older dikes fell from fissure walls so that the fissure changes width in amount corresponding to the collapsed sections (Fig. 2, D). Collapse is always proven where dikes widen abruptly downward, but where they widen upward it is generally impossible to determine whether it was caused by collapse of the upper part of an older dike or by renewed opening of a partly filled fissure.

The walls of most fissures show a surprisingly small amount of convergence or divergence that can be related to rotation of adjacent blocks during the growth of the fissure. This is probably due to the considerable depths of fissures in comparison to the limited vertical range over which they are generally exposed rather than to nonrotational separation of the walls. The deepest exposure seen, in a washout of the Burlingame ditch 3 miles southeast of Touchet, shows a fissure (Pl. 3, fig. 1) less than 12 inches thick extending through a vertical range of 120 feet. Most fissures that penetrate only a few feet narrow and pinch out rapidly. The lower portions of fissures either pinch out downward or end abruptly along bedding planes and in disturbed zones of the enclosing deposit. Many fissures narrow upward, but few pinch out in that direction; at the top most of them end very abruptly at the bases of current-bedded layers.

Some of the openings that received sedimentary fillings can scarcely be termed fissures, for they are of rounded, lenticular, and variously shaped cross section. Some are evidently sections of tunnel-like openings that carried underground streams; others are more nearly equidimensional cavities that were filled with sediment from connecting fissures.

Fault displacements.—Slight vertical displacements are common along fissures and in some cases involve movement of 1 to 10 feet (Pl. 5; Pl. 6, fig. 2). Displacements of dikes and fissures along low-angle and horizontal faults are also common (Pl. 4, fig. 2; Pl. 6, fig. 2); some are of only a few inches, a few range up to 8 feet, and one exceptional example in a highway cut 2 miles east of Wallula shows cumulative horizontal displacement of 22 feet on five faults.

STRUCTURAL DETAILS WITHIN DIKES

Clay layers.—Fissure walls are coated with a layer of clay which is generally less than 1 millimeter thick. The layers sharply delineate the margins of the dikes, for they are almost invariably of lighter color and finer grain. Most layers are finely laminated and, in addition, commonly have a coarser but still minute stratification that indicates periodic growth. In some layers silt and fine sand is interlaminated with clay. Most layers are, individually, of fairly uniform thickness, but some are vertically ridged.

Many clay layers occur as transverse layers in the dikes where they apparently mark interruptions of the filling processes. In stratified dike sediment they parallel the stratification and seem to be nothing more than fine-grained laminae produced by the stratifying process, but, in both stratified and unstratified dikes, some of the transverse layers bend upward and follow the fissure walls, and a few turn downward along the dike margins. Fissures only 1 or 2 millimeters wide are generally filled with clay, and some nearly horizontal sections of dikes, as much as 10 millimeters thick, are filled with finely laminated clay, silt, and very fine sand.

Degree of sorting and stratification.—The dike materials between the marginal clay layers show various degrees of stratification and sorting (Fig. 2). Unstratified sediment predominates and ranges from very well sorted material, mostly sand, to wholly unsorted mixtures of silt, sand, and gravel. Stratification is very common, and in some localities fully 50 per cent of the dike sediment is well stratified. Most dikes that can be observed for a depth of more than 10 feet show much heterogeneity of structure and composition; sections of laminated fine sand may be succeeded, in order, by unsorted mixtures of coarse sand and silt, well sorted unstratified sand, gravel, and massive silt, or any other combination of structure and composition.

The usual stratification is one in which the layers, or laminae, dip at angles near 30°. They resemble foresets without the usual topset and bottomset components and extend from the footwall of the fissure obliquely downward and across to the hanging wall (Fig. 2; Pl. 4, fig. 1). Therefore, in dike sections with dips of only 30°–60°, the foresets are long and meet the fissure walls at very acute angles, and in dike sections with dips of about 30° they parallel the fissure walls (Fig. 2, B). In irregular or zigzag dikes that reverse the direction of their dip, the dip of the foresets reverses correspondingly (Fig. 2, A). Most of the foresets dip nearly normal to the fissures, but in some dikes the dip is very oblique to the fissure and in exceptional cases it even parallels the strike of the fissures so that the foreset beds are narrow but of considerable length and depth. Rarely does the dip of the foresets exceed the maximum angle of repose, but several cases of high dip (55°) have been observed (Pl. 4, fig. 1). The high dips may be caused by contemporaneous deforma-

tion of enclosing deposits, which was commonly sufficient to rotate dikes more than 20°, and by lateral compression of dikes. Horizontal stratification in dikes is rare. Stratification of saucer- or cup-shaped cross section (Fig. 2, B; Pl. 4, fig. 2) is more abundant. Dike sections of low dip contain fine sand and silt that is well stratified and, in some cases, minutely cross-bedded. An exceptional stratification, limited to rounded and irregular sand-filled cavities, is concentric (Fig. 2, I). Crumpling of stratified dike sediments (Fig. 2, J) by differential movement of fissure walls is rare.

MODES OF FISSURE FILLING

SOURCE OF MATERIALS

The clastic dikes are unusual in that they show no evidence of plastic or water-charged sediment injected from below in the manner advocated by Diller (1890), Newsom (1903), and others for most previously described clastic dikes. Evidences that most fissures reached the surface of the ground and were filled from above are legion. In many places the dike sediments are so unlike the adjacent and underlying deposits that injection from below is ruled out. On the other hand the sediments immediately above the dikes are generally similar to those in the dikes (Pl. 3, fig. 2). Many dikes are traceable to overlying current-bedded sand, and, in some cases, the final stages of filling and covering of the fissures can be read with some assurance in the stratified sand (Fig. 2, C). It is not uncommon to find fully half the dikes of an extensive exposure ending above at a bedding plane in otherwise massive sediments; some of these planes evidently mark surfaces along which sediments were swept into open fissures, and others are surfaces of contemporaneous erosion that have truncated older dikes. The dike sediments show various features, such as stratification, sorting, alternation of silt, sand, and gravel increments, and collapsed wall materials which admit no other explanation than upbuilding processes.

Filling was accomplished in at least four general ways: (1) Most of the sediment was brought by streams, lake currents, and waves moving across the tops of fissures, (2) some collapsed from fissure walls and poured in from unconsolidated surficial layers, (3) a small amount was carried by underground currents, and (4) films of mud were spread on fissure walls in several ways. Wind-borne sediment is to be expected but is not proven. Most fissures were filled by two or more processes operating either together or alternately.

MOVING WATER ABOVE THE FISSURES

In this process, which probably accounts for more than 75 per cent of the dike material, the shifting sediment of stream and lake beds was trapped by water-filled fissures that reached to the surface of the earth. The sediment cascaded down the footwalls of the inclined fissures and, reaching the top of the upbuilding dike, assumed an angle of repose, thus forming the common foreset bedding oblique to the fissure walls (Fig. 2; Pl. 4, fig. 1). Thus the stratification in the dikes reflects, in many cases, a fluctuation of currents and materials taking place in water many feet above the surface of deposition; it is truly stratification by remote control. The common

development of foresets with dips higher than 25° shows that the filling was very gradual because, as can easily be demonstrated experimentally, rapid increment of sediment sets up density currents which carry the sediments rapidly across the fissures and develop low-angle, horizontal, or troughlike layers.

Much of the sediment came in fairly equal amounts for considerable distances along the fissures so that the dip of the foresets is mostly crosswise of the fissure, but in some places the entry of sediments at localized points caused the foresets to dip obliquely to the fissure walls and even advance lengthwise of the fissure. A control of the foresets by the dip of fissure is admirably shown by zigzag or curved dikes with reversed dips, for in these the foresets reverse in conformity with the dip of the fissure (Fig. 2, A). Few dikes are so nearly vertical as to avoid control of foresetting by the footwall, but occasionally short vertical sections show interruption of the normal types of foresets and a replacement of them with troughlike, horizontal, and ridged stratification caused by sediments sliding down both walls, falling equally on the entire width, or streaming down the middle from a projection or angle in the fissure (Fig. 2, B, C, D).

Probably some of the dike stratification was induced by the differential rate of sinking acting upon small increments of unsorted sediments, including those falling from the fissure walls. The coarse grains, sinking rapidly, would be followed by the finer grains, and successive additions would produce marked transitions of coarse and fine material in the dike sediments. Stratification by differential rate of sinking is strongly indicated by successive layers that grade upward from coarse sand to silt (Fig. 2, B), but in most stratified dikes the two processes cannot be differentiated. Differential rate of sinking, under some conditions, would nullify the stratifying effect of currents above the fissure and provide unstratified dike sediments.

It is commonly impossible to determine, from the dike materials and structures alone, whether streams, lake currents, waves, or winds brought the sediment into the fissures. The emphasis given here to moving water is based upon the abundance of gravel and coarse sand in the sorted or stratified dike sediments, the lack of aeolian sediments in the proglacial beds, and the common origin of dikes in overlying current-bedded layers composed of materials like those in the dikes.

COLLAPSE AND INFALL

The collapse of material from fissure walls is necessarily a minor process of fissure filling because the material it adds to a fissure is matched by a corresponding increase in width of the unfilled portion, and the volume of sediment required for filling a fissure remains unchanged. A surprisingly small amount of sediment fell from fissure walls even though most fissures opened beside older dikes; hanging walls that are nearly horizontal or of low dip seldom crumbled. Nevertheless, close examination of a number of dikes reveals abrupt changes in fissure width and irregularities on opposite sides of the dike that do not correspond, so that, if the dike material were removed and the fissure closed again, some openings would remain. This is particularly evident where older dikes formed the fissure walls, for their collapse is marked by an abrupt ending of the part that remains and a corresponding widening of the new dike. In a number of places the lower parts of dikes have col-

lapsed into the fissure, leaving the remaining part above as an unsupported layer of unconsolidated sediment and separated from still older dikes by laminated clay layers but nevertheless adhering to the fissure wall (Fig. 2, D, E).

Collapsed material in the dikes is proven by tumbled fragments of the clay seams that line fissure walls, and it is suggested by unsorted mixtures of silt, sand, and gravel. Along the north side of the Clearwater River, Idaho, fissures originating in and beneath deposits of hillside rubble and talus penetrate underlying silts and are filled with talus and rubble that evidently collapsed from above (Pl. 3, fig. 2).

Probably loose sediments collapsed or poured into fissures that opened beneath newly deposited sediments in stream channels, ponds, and lakes. A process of this nature should be indicated by disturbance, particularly downward bending, of strata at the top of the dike and by a flaring, funnel-like top of the fissure. Similar structures would be present if the squeezing and flowage modes of filling suggested by Jenkins (1925a; 1925b) and Flint (1938, p. 500) were correct. Such structures have not been certainly recognized. Downward-bent strata at the tops of dikes have been seen, but they show the final stages of fissure filling by moving water whereby the fissure was gradually transformed into a troughlike channel and finally buried by horizontal strata (Fig. 2, C).

CURRENTS WITHIN THE FISSURES

Occasional small fissures and sections of large fissures are of such low dip that sediments could not enter or pass through them by means of the simple gravity process described above; nevertheless they are filled with clay, silt, and sand. The silt and fine sand has excellent stratification parallel to fissure walls and occasionally minute cross-bedding, showing that currents moved down and along the fissures and redistributed sediments that slid down adjacent steep walls.

The existence of underground currents of sufficient strength to move sand and silt seems to be a logical adjunct of the scabland-Touchet episode. The changing levels of Lake Lewis, especially of the recessional phases, should have produced strong underground drainage, and the fissures would be avenues of easy flowage. Many of the coarse scabland deposits blocked the mouths of canyons along the Snake River and must have been permeated with strong currents entering and leaving the impounded areas behind the dams. Water moving downward in the fissures and entering the porous deposits is to be expected also. In addition, the fissure sections of low dip are normally constricted portions because of the predominantly horizontal separation of the fissure walls and so must have concentrated and speeded up underground currents, especially after other more accessible portions were filled by gravity.

Small-scale slumping and crumpling resulting from plastic flow has been seen on several dike sections of low dip. Density currents developed by sediments coursing down the footwalls of steep sections may have carried sediments across horizontal sections, but their records have not been recognized in the dike sediments.

A good example of current transportation within fissures is found in a large complex dike in the Asotin sand pits. The steep sections of the dike show foreset bedding and other structures characteristic of gravity filling, but the many dike sections

that dip less than 20° are mostly stratified parallel to the fissure walls, and here and there the sediments are disturbed by plastic flow. The sand deposit blocked the mouth of Asotin Creek canyon, and strong inflow and outflow of water is demonstrated by foreset bedding in large channel fills that traverse the sand deposits.

WIND ABOVE THE FISSURES

It is to be expected that strong winds would be present in southeastern Washington during glacial stages, as they were in other regions near the fronts of continental glaciers. The large barren exposures of unconsolidated silt and sand exposed during recessional stages of Lake Lewis should have presented conditions ideal for the filling of fissures by sediments shifted about by the wind. Much of the dike sediment shows sorting and grain sizes compatible with aeolian deposits; the clay layers on fissure walls and much of the common foreset stratification in dike sediments could have been developed in sediments swept into fissures by wind.

However logical the aeolian process may be, there is little evidence to support it. Some of the lack of evidence lies in the fact that the usual diagnostic criteria of aeolian deposits cannot exist in sediments that move short distances on the surface and pass downward through water-filled fissures. Possibly some of the foreset stratification was formed in dry fissures by aeolian sediments. Fine sand and silt fed slowly into a dry fissure will develop an angle of repose comparable to that of water-laid sediment, and on moist dike surfaces will stand on slopes as steep as 55°, but coarser grains bound from wall to wall, concentrate against the hanging wall, and develop a low angle of repose. The latter case is rarely suggested by the dike sediment. Against the aeolian process may be listed also the lack of aeolian deposits in the beds that contain dikes and the considerable amount of dike material, such as gravel and well-sorted coarse sand, which could not have been swept into fissures by the wind. Therefore this process must be considered largely unsubstantiated.

ACCRETION OF CLAY AND SILT

The seams of clay and silt that line fissure walls and extend across the dikes probably were formed by several processes. Streams of muddy water that run down steep, fresh exposures of porous sediments spread discontinuous mud layers through evaporation and filtering. The continuity and generally uniform thickness of seams indicate that this process is inadequate. Continuous mud films are formed by dust that collects upon moist surfaces and by evaporation of ground water that carries interstitial clay from sedimentary deposits. These processes operate only above the water table. Periodic recession of Lake Lewis and seasonal fluctuation of the water table may have made the processes locally effective, especially on canyon and valley slopes, but there is no evident basis for the complex inferences that would result from an assumption that all clay seams, in this region of lake and stream deposition, formed above the water table. It is especially difficult to admit that the ground water, in the broad, low, lake floor areas of the southern basin region, dropped below all deep fissures that have clay seams and rose again in all that contain water-borne sediments. It is doubtful that streams of muddy water would follow hanging walls of low-angle fissures, or that deep fissures, a few millimeters

wide, would permit appreciable evaporation of ground water or dust accumulation. The evaporation process is contradicted also by laminated clay layers that include laminae of silt and fine sand, for the transfer of such materials through clay laminae to the surface of evaporation seems impossible.

The writer believes that much of the clay was deposited from muddy water that stood in the fissures above the upbuilding dike. Probably much surface water entered the ground water stream through walls of fissures, thus filtering out the mud and leaving it upon fissure walls and surfaces of interrupted deposition within the dikes. This process accounts for the continuity, uniformity of thickness, and lamination of clay seams, and for their presence in tiny cracks and deep narrow fissures. This conclusion is supported by clay and silt laminae of wall seams that are continuous with laminae in stratified, water-laid, dike sediments; on some nearly horizontal dike segments, clay laminae in current-bedded sediment turn upward along adjacent steep walls.

In many cases it was impossible to distinguish between very thin clay dikes and wall seams of prominent dikes. Moreover, some transverse seams turn downward along fissure walls. These conditions suggest that slight additional opening of fissures, in some cases along incomplete dikes, caused fine-grained sediments to fill marginal cracks. Probably some seams, or very thin dikes, were formed in shrinkage cracks along dikes during postglacial compaction of the deposits.

ORIGIN OF FISSURES

PERTINENT OBSERVATIONS

The processes involved in the origin of fissures are more difficult to analyze than the modes of fissure filling, for they must be based upon less tangible criteria. This situation is very evident in the literature on clastic dikes. Some previously described clastic dike fissures have been rather definitely related to disruption of lithified rocks by orogenic forces, and others are known to be solution channels. In some cases the forcible injection of plastic dike materials has been given emphasis as an aid to or a cause of fissure growth. Development of fissures by earthquakes has received much attention, especially by Diller (1890), and this concept was applied to Pleistocene fissures near Lewiston and Touchet by Jenkins (1925a). There are no tangible criteria for identifying ancient earthquake fissures, and this interpretation depends largely upon the reasoning that earthquake fissures have formed in recent times, that sand and water have been erupted through them, and, therefore, they offer a simple, and perhaps logical, means of accounting for both dikes and fissures.

The usual explanations of fissure genesis cannot be applied to the Pleistocene fissures of the Columbia Basin. The inadequacies of solution and forcible injection processes are obvious. Though there may have been some regional changes of level, minor faulting, and earthquakes when the dikes were formed, there was certainly no marked local disturbance of the rocks beneath the Pleistocene deposits except in landslide areas. It is not certain that earthquakes did not in some way assist or speed fissure growth, but they were at best only incidental. Any interpretation of the fissures must take into account several limiting conditions. The

compound dikes show that fissures either opened repeatedly or opened slowly and continuously and were intermittently filled. The small amount of collapse shown by the fissure walls, even though those walls were made of unconsolidated sediments and newly formed dikes and, in places, had horizontal overhanging surfaces, suggests a gentle process of fissure growth. While it is perhaps conceivable that 80 dikes in one fissure represent 80 earthquakes, it is difficult to believe that earthquakes can be harmonized with the slight disturbance of fissure walls unless the sediments were frozen or in some other way made more resistant to collapse than they are now. Composition and grain size of sediments are not determining factors, for fissures were formed in materials ranging from silt to coarse gravel. The abundance of dikes in clean gravel and coarse sand deposits discounts the importance of settling and cracking resulting from ordinary compaction. Moreover, if compaction could produce such numerous fissures, fissures and clastic dikes should be among the most commonplace geological features in deposits of all ages.

It is most significant that of the various Pleistocene and recent sedimentary deposits of the region only the proglacial deposits contain abundant dikes, though in exceptional cases the dikes penetrate underlying basalt, and in near-by regions sedimentary dikes of another kind are present in Columbia River basalts. Moreover, the formation of fissures, while a long-continued process, was limited to the times at which the proglacial deposits accumulated. These conditions are strikingly evident where the dike-bearing deposits lie in contact with similar deposits of other ages. In Alkali Flat Creek canyon, for example, a postglacial canyon fill of silt was deposited upon and against erosional remnants of an older fill of Touchet silts, and both deposits are now trenched by recent gullies. In the excellent exposures of the gully walls no dikes were found in the postglacial silt, but each Touchet remnant is intricately divided by numerous clastic dikes. Likewise other fluvial and lacustrine sediments of Pleistocene age but older than the glacial stages have few if any dikes even though adjacent to dike-bearing deposits. In both Clarkston and scabland-Touchet deposits the fissures are lacking in well-washed stream gravels of major streams but are abundant in erratic-bearing deposits that include lake or slack-water sediments. All these observations indicate that the fissures were derived from some condition that was not entirely dependent on composition and grain size of sediments, recurrent but each time long-continued, not sudden or violent in its operation, unique to proglacial times, and most pronounced in slack-water deposits.

Proglacial restriction alone is sufficient to eliminate several processes as primary factors of fissure formation, for we must find a process, or processes, that came and went with the glacial episodes. Earthquakes and diastrophic phenomena in general, compaction, and plastic deformation by gravity sliding are thus eliminated or relegated to minor roles. It is believed that only processes related to the ice of glacial times can meet the requirements.

MELTING OF BURIED ICE

The thesis presented here is that most fissures in the Pleistocene deposits of the Columbia Basin are related to melting of buried ice and frozen sediments which

caused uneven settling and sliding of the sediments above and, in some cases, left variously shaped cavities.

Deformation of proglacial sediments through melting of buried ice is widely recorded in geological literature. In most cases the deformation occurs in glaciated regions and is attributed to glacier ice, but glacier ice, except as bergs, was not available in the southern Columbia Basin region. However, frozen ground and ground ice are present even today in large areas of Siberia, Alaska, and Yukon territory. They have been described by Leffingwell (1919), Cressey (1939), Taber (1943), and many others. Much of the ice occurs as distinct layers, lenses, and wedgelike veins of pure ice, and much is frozen muck and silt in which the water content ranges as high as 80 per cent of the volume. Ice and frozen ground reach depths of more than 300 feet, and frozen and unfrozen zones alternate vertically. Surficial ridges, troughs, and basins have been produced by subterranean melting. Various explanations of the source of ground ice have been advanced. Some hold that surficial ice, from lakes, streams, and spring waters, was buried by sediments; others believe that the ice was formed beneath the surface. Taber (1943) presents much evidence to show that ground ice in Alaska has been formed by segregation during progressive freezing from the surface downward.

It is impossible to determine, at present, the degree to which ground ice and frozen ground like that of present-day Alaska were developed in the Columbia Basin during glacial times, but there are important differences in that the Columbia Basin deposits are almost entirely lacking in vegetal material; they were formed in and near a large oscillating lake, and deep fissures remained free of ice, whereas the Alaskan and Siberian deposits contain much peat and intermixed vegetal material, no large lake was present, and nearly all openings, even small contraction cracks, are filled with ice. Though Taber (1943, p. 1527) argues that burial of surface ice by alluvium is difficult because the ice tends to float away, it nevertheless seems that conditions would be especially favorable for this process within the areas periodically flooded by Lake Lewis, or by local ice jams in scabland tracts, because the periodic recessions of the water should ground large quantities of lake and river ice, as well as bergs of glacier ice, where they might readily be buried by the small streams that reworked the surficial sediments during recessional stages. The absence of fissures and dikes from the well-sorted gravels deposited by the main current of Snake River probably means that fissures were not formed where there was sufficient current and depth of water to carry the ice away. At any rate, the abundance of ground ice in cold regions today, the evident abundance of floating ice in the Lake Lewis area, and the sedimentary conditions favorable to burial suggest strongly that subterranean ice, perhaps of more than one type, would be the normal result of proglacial climates in the Columbia Basin.

The proglacial deposits of the Columbia Basin show certain features that tend to confirm the inferences based on environmental conditions. The disturbed undulatory bedding, kettlelike sags, and small high-angle faults of the Touchet beds are especially suggestive of buried ice, for they indicate a process of irregular settling that cannot be accounted for by simple compaction. There is some reason to attribute the fissures to the same process that caused the small-scale contemporaneous

deformation. High-angle faults generally coincide with dikes (Pl. 5; Pl. 6, fig. 2). Likewise abundant dikes are generally associated with undulatory bedding and sags. This may be in part due to the frequency of both disturbed beds and dikes, but it is very noticeable that little-disturbed beds, such as those in the Burlingame canyon (Pl. 3, fig. 1) and Tammany Creek (Pl. 1, fig. 2), have only occasional dikes, whereas similar but disturbed strata near-by contain hundreds of dikes and fissures. This relationship seems to be maintained even within single exposures that show both disturbed and undisturbed bedding, and in a number of these there is ample reason to believe that the warping, faulting, and fissuring are all related to the melting layers, lenses, or blocks of ice and to thawing of frozen water-charged strata.

The downwarped and faulted sags of the Touchet beds record the gradual disappearance at depth of some object or material while sediments were being deposited on the surface. As the sags were seen only in cross section it was not possible to determine whether they are nearly equidimensional structures such as would be formed above melting ice blocks, or linear structures like those formed above ice wedges in the ground-ice regions of Alaska (Taber, 1943, p. 1514). In the bottoms of some of the sags are tumbled blocks of sediment surrounded by a maze of dike-filled fissures; in other places fissures converge downward toward the centers of the sags (Pl. 5, fig. 2) or end abruptly in disturbed strata in the bottoms of the sags. In some places downwarping and fissuring were noted in and above structures that are partly of the cut and fill type, suggesting that streams in erosional channels, and perhaps ponds in kettles, were periodically frozen and buried and, upon melting, caused deformation of sediments that were swept into the depressions.

Evidence of layers and lenses of ice, of greater horizontal dimensions than the pieces that formed the sags, were noted at several places. The former location of the ice is generally indicated by dikes that end abruptly below along the same bedding plane or along a zone of disturbed stratification and by warped and faulted beds lying on less deformed beds. In a few cases settling by compaction of underlying sediments cannot be ruled out, but in others the character of underlying materials definitely rules out compaction as an effective factor. On lower Tammany Creek near Lewiston, for example, fissured Clarkston sandstones lie upon a blue-gray lake clay, thus suggesting that compaction of clay was involved in the fissuring even though the small lake represented by the clay would have been an ideal place for the burial of lake ice. Nearby are Clarkston gravels and coarse scabland gravels which lie upon basalt bedrock; appreciable compaction of gravels and basalt seems to be out of the question, but the gravels are cut by clastic dikes. Moreover, above the scabland gravel a blanket of Touchet silt (Pl. 1, fig. 2) is traversed by fissures that end abruptly at the eroded top of the gravel bar, and along the contact of gravel and silt are lenticular, sand-filled cavities formed by the disappearance of some material which, it seems, could only have been ice.

More direct evidence of buried ice is found in sand-filled cavities which are of most common occurrence in Touchet beds. One type (Fig. 2, I) is of rounded or irregular cross section, a few inches to several feet in diameter, and filled with sand which may exhibit either the inclined stratification so common in dikes or a concentric stratification around one or two centers, evidently the product of successive

additions around melting ice blocks. Small fissures through which the sand entered are visible above some of the former cavities. These structures may be seen in the road cuts 0.8 mile southeast of Kiona, in the road cuts north of the Clearwater dam, and in the silt above the Tammany Creek gravel bar (Pl. 1, fig. 2). A second type (Fig. 2, F) is lenticular, generally 1 to 20 feet long and less than 12 inches thick, and conformable with the bedding. They show that lenticular cavities, or horizontal fissures, opened between layers of Pleistocene sediments, a process that seems impossible to explain by any kind of settling or disturbance of the beds. The Touchet silt on the Tammany gravel bar contains several zones along which these sand lenses are common. An exposure (Pl. 1, fig. 2) shows three of these zones: one at the top of the gravel, one 2 feet above the gravel, and one 10 feet above. They resemble ordinary lenses of sand interbedded with the silt but they are dikes or, more properly, sills, which are later than overlying silt beds. Here and there between the lenses but at the same horizons are rounded or irregular bodies of sand that bear no resemblance to surficial deposits. Vertical fissures, filled with the same kind of gray sand as that in the lenses, originate in the upper part of the silt, and most of them either end in the roofs of sand lenses (Fig. 2, F) or turn to a horizontal position and continue for a few inches or a few feet between silt beds (Fig. 2, H). The prominent lens of gray sand in the center of the view lies in a downwarped pocket in a silt layer which, in turn, overlies a zone of sand lenses and may represent a sand-filled subterranean kettle. Near by (Fig. 2, H) the silt beds have sagged and parted along prominent bedding planes. These features apparently require the gradual disappearance of some material that intervened between beds of silt, and so the former presence of ice layers and their connection with fissuring is strongly indicated.

The nearly horizontal faults common in Touchet beds apparently are related to the same source of disturbance that caused high-angle faults, sags, and undulatory bedding, and in some cases movement on the low-angle faults was accompanied by faulting and fissuring of the beds above. Some of the movement may have been caused by slumping and sliding of water-saturated beds on gentle slopes, which Flint (1938, p. 499) suggested as a major cause of deformation in the Touchet. A curved fault (Pl. 5, fig. 1) suggests this interpretation for it is a gravity fault in beds that lie on the side of the Touchet valley. Other low-angle faults apparently cannot be explained as valley-side slumps. The deep highway cuts in Touchet beds 2 miles east of Wallula (Pl. 6, fig. 2) show a number of nearly horizontal offsets of older dikes, one displacement on five faults amounting to 22 feet. Other compound dikes come down to one or the other of two prominent fault surfaces and end abruptly. The failure of these latter dikes to continue below the faults can be explained by assuming a fault so large as to exclude the lower parts of the dike from this large exposure, but this demands a complex and highly improbable sequence of faulting and fissuring. More probably fissures were developed by marked tension and pulling apart of the strata above the faults when the large compound dike was offset. Similar indications are found in the great width of some fissures, such as that one-fourth mile east of Kiona, which has a width of 45 inches and is filled with 80 dikes.

These features indicate some process other than slight rotation and cracking above melting ice. The usual type of gravity sliding of plastic sediments on an inclined

lake bottom seems to be excluded not only by the essentially horizontal attitude of the strata but by the fact that the sediments were disrupted after they had become firm enough to form deep, clean-cut fissures. Here again the melting of buried ice layers and frozen water-charged sediments offers a logical explanation, for the melting would give rise to plastic, water-charged zones conducive to sliding; unequal melting and settling, or undulations in the zone of melting, would furnish the temporary inclinations necessary for repeated sliding, fissuring, and growth of compound dikes.

A puzzling phase of the entire dike problem is the explanation of the conditions that made the newly deposited sediments so resistant to collapse on the walls of deep fissures. Some of these sediments even today are so little consolidated that vertical cuts in them are nearly buried in their own debris within a few years. Nevertheless, fissures with vertical, overhanging, and even horizontal walls remained open while they were gradually filled with sediment, newly formed dikes adhered to fissure walls without collapse, and some lenticular horizontal cavities remained open until they were filled with sand from intersecting fissures. In view of the evidence that melting of ground ice caused deformation and fissuring of the proglacial deposits, it might seem reasonable to suppose that the sediments above zones of subterranean melting were rigidly frozen. This concept would explain the peculiar combination of sags and folds which require the elasticity of thawed sediments, and fissures and faults which suggest the rigidity of frozen sediments. Thawed ground beneath frozen ground has been observed in Siberia and Alaska; Taber (1943, p. 1509–1510) attributes it to the circulation of thawing ground water along permeable layers, and Tuck (1940, p. 1301) believes that surficial depressions, ridges, and hummocks near Fairbanks resulted from subsurface thawing, doming of overlying frozen sediments by hydrostatic pressure, and later collapse.

In view of the glacial climate and the abundance of ice when the sediments were deposited, the writer cannot deny that some of the sediments were frozen when fissures developed, but a thesis that fissures formed only in frozen sediments meets insurmountable obstacles. The fact that water stood in fissures while they were gradually filled with sediment seems to show conclusively that frozen sediments were not necessary to prevent collapse of fissure walls. Water-filled cracks in the frozen ground of Alaska result in ice dikes, not clastic dikes (Leffingwell, 1919; Tuck, 1940, p. 1301; Taber, 1943). Moreover, even if it were possible for water to remain unfrozen in fissures that traverse frozen ground it would often cause thawing of wall materials, especially along compound dikes where many successive openings were filled with water. Thus we are confronted with the difficulty of explaining how newly thawed sediments, which are normally very loose or even fluid, remained firmly in place upon most fissure walls. Therefore the writer believes that the proglacial sediments were not frozen when the fissures were formed except for minor frozen zones probably near the surface and possibly also adjacent to buried ice. The apparent strength of the wall materials is probably best explained by (1) the gentle operation of the fissuring stresses above melting ice, and (2) the firmness of the sediments which are generally sandy and lack the high clay and vegetal content so common in other proglacial deposits.

Other conditions related to ice and cold may also have been effective in forming

fissures, but these depend upon deeply frozen ground. Deformation of sediments by thrusting of shore ice and ice bergs have been described by many writers, and Hodge (1931) attributes some deformation to thrust by ice jams. Stresses of this nature probably could not extend to the depths required by the fissures unless the sediments were deeply frozen. Expansion and contraction with changes of temperature, such as advocated by Tuck (1940, p. 1301) for fissures and ice dikes and by Leffingwell (1919, p. 206–214) for the growth of ice wedges and veinlets, likewise require deeply frozen ground, and Taber (1943, p. 1521) states that tension cracks due to temperature changes do not extend to depths of more than 5 or 10 feet. The writer can find nothing in the descriptions of the common ice veins, or ice wedges, of Siberia and Alaska (Leffingwell, 1919; Taber, 1943) that seems applicable to the Columbia Basin fissures, for the veins are flaring, wedgelike masses of ice that generally deform adjacent sediments by lateral growth, whereas the Columbia Basin fissures are tension cracks with undeformed walls.

SUBTERRANEAN STREAM CHANNELS

This type of dike-filled openings has been recognized only in the scabland-Touchet deposits of the Lewiston Basin. They were formed as erosional channels, or tunnels, by underground currents and later filled with sand and gravel carried by the streams. Some of the tunnels are of rounded or irregular cross section and so not readily distinguished from the cavities left by melting of buried ice blocks. However, in the gravel bar that lies in the mouth of Tenmile canyon south of Lewiston a tunnel about 7 inches wide and 24 inches deep is filled with horizontally stratified sand and small pebbles (Fig. 2, G). The gravel bar, deposited by the Snake River while running more than 200 feet above its present bed, probably because of an ice jam in the main canyon, blocked the mouth of the canyon and so must have been subject to strong underground flowage.

LANDSLIDE FISSURES

Along the north side of the Lewiston Basin clastic dikes of Touchet age penetrate the older Clarkston deposits and in places even extend into the basalt bedrock. These are a notable exception to the general rule that dikes and fissures of one proglacial episode do not extend to older rock units. All these exceptions can be definitely related to landslides. Extensive landsliding, characteristic of the Columbia River lava in the Lewiston Basin region, has been described by Russell (1901). Most of the landslides affect the uppermost lava flows because they are underlain by a prominent interbasalt member of shale and sandstone. Where this interbasalt member passes beneath the 1325-foot contour, as it does in the synclinal downwarp of the Lewiston Basin, it was saturated with the waters of Lake Lewis and so yielded readily to cause slumping and fissuring of the overlying basalt flows and Pleistocene deposits as well.

Dikes and fissures of the above category are well exposed at the foot of the Lewiston grade. They traverse Touchet silts and in places enter the bedrock. They were discussed by Jenkins (1925a, p. 238–239) who stated that "The origin of these filled cracks may be in the settling of the terraces or more likely from the cracking

due to earthquakes." The terraces have settled, and earthquakes may have been an assisting factor, but the primary control by landslide conditions is evident in faulted blocks of Touchet beds and lava flows and in landslide topography. Similar conditions are repeated along the Genessee road, 2 miles east of Lewiston, where dikes of unconsolidated gray sand penetrate tilted blocks of Clarkston sandstone and enter the lava beneath. Dikes associated with landslides have been seen outside the Lewiston Basin only in a deep highway cut 5 miles northwest of Touchet. Here are dikes of silt and clay in a rubbly hillside deposit, presumably of Clarkston age, which lies upon considerably tilted slump blocks of Columbia River lava.

SUMMARY AND CONCLUSIONS

During the Clarkston stage a gravel fill along the Snake River caused local ponding of less vigorous tributaries. During the scabland-Touchet stage the water of Lake Lewis rose to a maximum altitude of 1325 feet above sea and inundated the southern part of the Columbia Basin. Two contemporaneous lithologic facies were produced in the lake area: (1) the scabland deposits of gravel and coarse sand which were deposited by major streams, and (2) the Touchet beds of silt and fine sand deposited by lake waters, sluggish streams, and small streams. Sedimentary features of the Touchet beds, as well as the absence of features characteristic of lakes with stationary level, indicate that Lake Lewis periodically advanced and retreated, causing a rhythmic alternation of lake and stream deposition. Floating ice carried silt and sand as well as large erratics into all areas invaded by the lake. The arm of Lake Lewis that extended eastward along the Snake River canyon was clogged by ice at times so that currents were concentrated between ice and canyon walls, thus producing high-level erosional features. Gravel and coarse sand were deposited on and near the eroded tracts, especially in the mouths of tributary canyons. Silt, fine sand, and erratics were spread rather generally over the area, especially in tributary canyons and on the sides of the Lewiston Basin. During recessional stages of the lake the river reworked the sediments of the canyon bottom and left deposits of well-sorted gravels with rounded stream-worn pebbles.

The clastic dikes were formed only under conditions prevailing during glacial stages and are restricted to proglacial deposits except in rare cases where they penetrate subjacent rocks. Conditions favored formation of dikes where silts, sands, and unsorted gravels were deposited in ponded water or sluggish streams, especially where periodic retreat of the lake permitted small streams, lake-shore currents, and waves to sort and redistribute the lake sediments. Strong deep currents, such as that of the Snake River during recessional stages of the lake, prevented the formation of dikes.

Fissures reached the surface of the earth and were filled from above. The dikes show a wide range of structures and materials resulting from processes operating above and within the fissures. Most dike sediments came from lake and stream currents that moved across open fissures; some probably were moved by wind. Within the fissures the trapped sediments, together with smaller amounts of collapsed wall materials, generally were carried down by gravity, though some were shifted by underground currents, and films of clay and silt adhered to fissure walls.

Earthquakes and other diastrophic phenomena, compaction, solution, and forcible injection of sediments are inadequate to account for the fissures. The occurrence of dikes in proglacial deposits, and structures of those deposits, indicate that melting of buried ice was largely responsible for fissuring and small-scale faulting and warping. Three ice processes are indicated: (1) uneven settling and cracking above melting ice and frozen sediments, (2) sliding on inclined zones of melting, and (3) formation of cavities by melting ice blocks and layers. A few fissures were formed by landslides and underground streams.

The interpretations advanced here are opposed to certain earlier concepts of Pleistocene history. The rhythmic bedding of Touchet beds, periodic or long-continued opening of fissures, gradual filling of fissures by lake and stream sediments, and burial of ice are some features that cannot be harmonized with a single great "Spokane flood" or a single advance and retreat of Lake Lewis. The dikes and fissures agree well with the rhythmically bedded Touchet deposits in suggesting repeated flooding of the Lake Lewis area but throw no light on the cause of the ponding.

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ABSTRACTS

Because the number of abstracts submitted is considerably reduced owing to cancellation of all meetings except the Fall meeting of Section E, we do not think publication of the usual Part 2 of the December Bulletin is justified.

THE GEOLOGICAL SOCIETY OF AMERICA

CENOZOIC WEATHERING PROFILES IN THE NORTHWEST*

BY VICTOR T. ALLEN, ROBERT L. NICHOLS, AND VERNON E. SCHEID

During the Cenozoic, northwestern United States was a land area except for a narrow coastal belt which was at times flooded. During this era there was intermittent deposition of continental sediments and pyroclastic rocks and intermittent extrusion of lava. Successive land surfaces of erosional and depositional origin were formed and at most places buried. Where the time interval was sufficiently long and where physiographic and other conditions were favorable weathering profiles were formed on these surfaces.

More than 12 profiles formed at different times and places have been recognized, and there are undoubtedly others. They were developed in every epoch of the Cenozoic on gneiss, granodiorite, pegmatites, peridotite, basalt, tuffs, gravels, and other rocks. They range in thickness from a few to 100 feet and were formed under temperate and subtropical conditions; kaolinite, halloysite, montmorillonite, beidellite-nontronite, gibbsite, diaspore, boehmite, limonite, garnierite, and other minerals and substances were developed in them.

Nearly all the high-alumina clay deposits in the Northwest as well as some nickel and iron deposits are genetically related to these profiles. Residual materials of the profiles comprise the Excelsior, Washington, and the Benson, Idaho, clay deposits, the high-alumina iron deposit of Washington County, Oregon, and the nickel deposit near Riddle, Oregon. The sedimentary high-alumina clay deposits near Castle Rock, Wenatchee, and Spokane, Washington, Troy and Moscow, Idaho, and also the Scappoose, Oregon, and Cle Elum, Washington, iron deposits that are partly residual and partly transported, were derived from them. Sedimentary clay at Molalla, Oregon, is interbedded with the products formed in three superimposed profiles.

RAINBOW ICE CAP, BEARTOOTH MOUNTAINS, MONTANA

BY ARTHUR BEVAN

During late Pleistocene time a part of the subsummit plateau in the northwestern portion of the Beartooth Mountains was covered by an ice cap. It is here named the Rainbow ice cap from the Rainbow Lakes on the Lake Plateau of the Mt. Douglas quadrangle.

The area of the ice cap is within the drainage of two major north-flowing tributaries of Yellowstone River—Boulder and Stillwater rivers—mainly in Sweetgrass County, about 20 miles northwest of the northeast entrance to Yellowstone National Park at Cooke, Montana. The altitude of the glaciated plateau ranges generally from 9000 to 10,000 feet. A few peaks rising above 10,000 feet may have been nunataks. Large distributaries of the ice cap which spilled down the several headwater canyons tributary to Boulder and Stillwater valleys greatly augmented the glaciers in those valleys.

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The relations of the Rainbow ice cap to the ice caps or ice fields on other parts of the subsummit plateau to the southeast, and to the huge glaciers directly north of Yellowstone Park, are yet unknown. It may be that this ice cap, though a functioning unit, was a part of the extensive ice field on the southwest slope of the Beartooth Range. (Geological Society project.)

FACIES VARIATIONS IN THE MIDDLE ORDOVICIAN OF THE APPALACHIAN VALLEY IN VIRGINIA*

BY BYRON N. COOPER

Regional studies of the Middle Ordovician in the Appalachian Valley in Virginia show that some of the familiar units, generally regarded of different ages, are laterally continuous and equivalent. These profound facies variations necessitate a reclassification of the Middle Ordovician in Virginia.

Two examples of major variation in facies are especially noteworthy. In the limestone belt along the northwest base of Clinch Mountain, 300 feet of "Holston" marble at Luttrell, Tennessee, grades northeastward into a succession including the Blackford, Five Oaks, Lincolnshire, Ward Cove, and Peery formations which are typically developed in Tazewell County, Virginia. The stratigraphic range of "Holston" marble beds, from Blackford to Wardell ("upper Ottosee"), clearly indicates the megafacies character of the "Holston."

In northern Virginia, the thick body of black "Athens" limestone near Harrisonburg grades northeastward into a thinner succession mainly of cobbly, buff-weathering limestone. The Chambersburg of the type section in Pennsylvania and along Tumbling Run, near Strasburg, Virginia, is a mixture of these two facies. Equivalency of the "Athens" and "Chambersburg" in northern Virginia is well supported by fossils.

In contrast, a few key zones, such as the *Reuschella edsoni* zone directly below the Martinsburg in the Shenandoah Valley and the Witten limestone beneath the Moccasin in southwestern Virginia, are persistent over large areas.

PERIGLACIAL CLIMATES AND PERIGLACIAL PHENOMENA

BY STEPHEN TABER

In the reconstruction of Pleistocene periglacial climates several modes of approach are available, the most important being: (1) study of climatic conditions in the vicinity of existing ice caps; (2) study of the present atmospheric circulation and climatic belts as a basis for estimating the effects of the large Pleistocene ice sheets; (3) study of the physical effects of periglacial climates; and (4) study of the effects of glaciation on the distribution of plants and animals. These studies indicate that the planetary system of winds in North America was slightly modified and accentuated rather than obliterated by a glacial anticyclone.

Certain soil structures peculiar to periglacial zones, such as involutions and associated clastic dikes, are due to thawing, which followed the deep freezing of fine-grained soils and resulting formation of segregated ice. This type of evidence indicates that near the extreme southern limit of glaciation the annual mean temperature was close to 26°F. The difference between present temperatures and Ice Age temperatures is greatest in the vicinity of the ice front and least in the vicinity of the equator.

Periglacial climates were very different in different localities. In North America the greatest contrast existed between periglacial climates in the south and in the far north. The climate was characterized in the south by moderately low temperatures, heavy precipitation, and strong winds; in the north by lower temperatures, light precipitation, and light winds.

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NEW TEXAS LOWER CRETACEOUS *GRYPHAEA* (OSTREIDAE)

BY W. S. ADKINS

A new species of *Odontogryphaea* is described. It is an upper Washita marker abundant in the Grayson marl and has generally been erroneously called *Gryphaea mucronata* Gabb. The latter species was described from the Fredericksburg group at Arivechi, Sonora, and is a well-known Fredericksburg marker, generally and erroneously called *Gryphaea marcoui* Hill and Vaughan. *Gryphaea* Lamarck 1801 contained six valid species, and the earliest valid subsequent designation from among these appears to have been that of *G. arcuata*, a Liassic species, by Chenu, 1858. The present species appears to be referable to the Lower Cretaceous genus *Odontogryphaea* v. Ihering 1903.

SILICIFIED PLANT REMAINS FROM WESTERN NORTH AMERICA

BY CHESTER A. ARNOLD

The opportunities for paleobotanical research among the silicified plant remains of the western States have been widely neglected. Some reasons for this neglect are the scarcity of paleobotanists interested in plant anatomy, the feeling that the necessary effort would not be justified by the results obtained, and the lack of many apparent connections with economic and stratigraphic problems. Most copiously represented are the silicified trunks and limbs of conifers and hardwoods, followed in order of abundance by palms, cycadeoids, and ferns. Silicified trunks of the arborescent lycopod *Lepidodendron* occur in the Lower Pennsylvanian of central Colorado. Largely through the efforts of Lester F. Ward and George R. Wieland the cycadeoids have been comprehensively studied, and several hardwoods, conifers, and palms have been described by various authors in shorter contributions. Few ferns have been described, the list consisting of *Osmundites skidegatensis* (Penhallow) from the Lower Cretaceous of Queen Charlotte Island, *O. walkeri* (Daugherty) from the Chinle formation of Arizona, and three species of *Tempskya* (Seward, and Read and Brown) from the Colorado group. The author has assembled five new ferns which will shortly be described in detail. They are a new species of *Osmundites* from the Tertiary of central Oregon, a new genus of the Aspleniaceae from the Green River of Wyoming, a tree-fern belonging to the Cyatheaceae from the late Eocene of Texas, and two species of *Tempskya* from the Morrison formation of Montana and Wyoming.

LOWER PENNSYLVANIAN SECTION EXPOSED IN THE ELK MOUNTAIN FAULT ZONE, GUNNISON COUNTY, COLORADO

BY CHARLES FERNANDO BASSETT

The 550-foot section of fossiliferous lower Pennsylvanian strata in an isolated fault block between Mt. Belleview and Snowmass Mountain is described, and its stratigraphic position suggested. A list of fossils from the McCoy and Deep Creek areas is included for comparison. Some details of the structure are added and discussed. Reference is made to the character of the Maroon formation as exposed in this zone.

HYDROZOAN JELLYFISH FROM THE LOWER CRETACEOUS OF TEXAS

BY KENNETH E. CASTER

Medusoid molds and casts in the Pawpaw formation (Washita series of the Comanchian) of Denton County, Texas, appear to be Trachyline Hydromedusae. The discovery marks the first finding of "jellyfishes" in American Mesozoic and offers paleontologic data on a group hitherto but little known in geology. A new family, genus, and species are required for the organisms. All close comparisons are with European materials, many of which have been referred to the higher divisions of the Scyphozoa. Should the present views prove to be correct, rather sweeping revision of medusoid Problematica must be undertaken.

NEW SYNZIPHOSUROID MEROSTOME FROM THE LOWER ORDOVICIAN

BY KENNETH E. CASTER

The Lower Ordovician "sink-hole" fauna on the foundation site of the Douglas Dam on the Tennessee River (Laurence, 1944) contains an abundance of carapaces of a small and very primitive

synziphosuroid merostome. This marks the first pre-Silurian record for organisms of this habitus, and, if it is to be contained in the Synziphosura, it will require critical modification of our present concept of the suborder. Family, genus, and species are assuredly new. The material is unique in the possession of 11–12 annular abdominal segments between the four (or five) segments of the "thoracic" region and the terminal "telson." An extraordinary and anomalous feature is the presence of a large ventral thoracic buckler analogous to the familar dorsal structure in *Limulus*. A new phyllocarid is associated with the merostomes.

WAS *RHIZOSTOMITES ADMIRANDUS* HAECKEL OF SOLNHOFEN A TRUE JELLYFISH?

BY KENNETH E. CASTER

The traditional assignment of the famous *Rhizostomites admirandus* Haeckel of the Jurassic (Solnhofen) deposits of Bavaria requires re-examination. The criteria of its hitherto unquestioned reference to the Rhizostomeae, highest order of the true jellyfish (Scyphozoa), are most highly questionable. Virtually all critical rhizostome attributes are unknown in the fossil. In contrast, in the fossil, in the strict sense of the name, and especially in associated remains of other assignment but probable conspecificity, most of the critical characters of the hydrozoan medusae can be ascertained. It is at least morphologically plausible that *Rhizostomites admirandus* Haeckel is a Hydromedusa having the aspect of certain modern Trachylina. It seems highly to be desired that all fossil records of the Scyphozoa be restudied and that paleontologic minds be open to the possibility of much more general preservation of the Hydromedusae than is currently supposed.

EARLY CENOZOIC SEDIMENTS IN THE APPALACHIAN REGION*

BY P. E. CLOUD, JR., AND R. W. BROWN

At Anniston, Alabama, 70 miles north of the coastal plain, fossil plants tentatively identified as *Asplenium eoligniticum, Dryophyllum tennesseensis, Banksia tenuifolia, Cinnamomum tennesseensis*, etc., occur in clay in an ancient sink in carbonate rocks of the upper Conasauga formation (Upper Cambrian). They were discovered in a pit 280 feet north of the intersection of the Eulaton and Morrisville roads in the NE $\frac{1}{4}$ sec. 11, T.16 S., R. 7 E. Similar occurrences were found near by, and additional collections were made. The plants include both Eocene and Cretaceous types, suggesting a Paleocene (Midway) or an early Eocene (Wilcox) age. No marine organisms were seen.

Subsequently *Banksia tenuifolia* was found in clay on the dump of a bauxite mine near Cedartown, Georgia, and fragmentary plants generally similar to those at Anniston in two clay deposits filling ancient sinks in dolomite of Lower Ordovician age near Fort Payne, Alabama.

It is hoped later to discuss the bearing of these occurrences on the age and origin of Appalachian bauxite. For the present, attention is called to them because of their significance to the geomorphic history of the region. The quartzite ridges northeast of Anniston, some with water-worn quartz pebbles on their summits, rise 1000 feet above the clay pits there; the other plant localities also lie in valleys well below the present ridge crests. Is present Appalachian topography exhumed, and how closely does it reflect that of early Cenozoic time?

FURTHER NOTES ON THE PALEOZOIC STRATIGRAPHY AND STRUCTURE OF NORTHEAST ALABAMA*

BY P. E. CLOUD, JR., K. M. WAAGÉ, AND N. M. DENSON

Information resulting from continued mapping of residual materials in southern Cherokee County has improved structural interpretations and revealed errors in Cloud and Waagé's abstract (1942).

Certain lobes of Rome arenites, formerly thought to be lenses in argillaceous beds carrying a Rutledge fauna, are now believed by Cloud to owe their unusual position to tangential marginal incision of the overriding block of Hayes' Coosa thrust by subsequent high-angle faulting; erosion has isolated them as separate fault blocks in a Conasauga terrane.

Present interpretation regards the patches of Frog Mountain sandstone studied as erosional outliers in a synclinal basin, resting unconformably on strata ranging in age from the lower Chepulte-

*Presented with the permission of the Director, Geological Survey, U. S. Department of the Interior.

pec dolomite to the Athens shale. A combination of local faulting and unconventional topographic relationships earlier led to the untenable hypothesis of graben structure. The relationships of these truncating Devonian sandstones appear to indicate post-Middle Ordovician and pre-Onondagan warping in northeastern Alabama.

Recalculations from better geometric evidence show the five units of the Cambrian and Ordovician Knox group to have a total thickness near 2500 feet.

The residual elements of the Shady dolomite, as mapped by us, compare closely with those in and near Shady Valley, Tennessee, and this dolomite is now known at a number of localities in the area.

Bauxite occurs locally as sink fillings in the carbonate rocks of the Conasauga and Shady formations as well as those of the Knox group.

BRYOZOA FROM THE PERMIAN OF WEST TEXAS

BY G. E. CONDRA AND M. K. ELIAS

Description and analysis of these Permian Bryozoa are based on the collection by P. B. and R. E. King, and Charles Schuchert, submitted to the Nebraska Geological Survey for study and description, and on a subsequent addition of the material collected by G. E. Condra and C. O. Dunbar and of the material etched from the Leonard limestone under the supervision of Dunbar. Only a few of the Bryozoa came from the Wolfcamp; most are from the Leonard and Word. We redescribe and revise a few species of fenestrate Bryozoa, previously described from the same area by Girty, and we describe many additional delicate, well-preserved forms, obtained by dissolution of selected limestone specimens in which the silicification, which is so common in Texas Permian limestones, did not obliterate finer structural features and forms. These forms include many species of *Pinnatopora*, *Diploporaria*, *Septopora*, *Fenestella*, *Polypora*, *Streblotrypa*, *Rhombocladia*, and others. From all collections 15 genera and 47 species and varieties are described.

STUDY AND REVISION OF GENUS *THAMNISCUS*

BY G. E. CONDRA AND M. K. ELIAS

The genus *Thamniscus*, as originally defined on Permian forms and subsequently expanded over species collected from older rocks, ranges from Silurian to the end of Paleozoic but is abundant only in the Permian of Western Europe, North America, India, and Timor. Material from the Midcontinent and India shows constant dense encrusting phytomorphic tissue on the reverse and, particularly, at the base of the zoaria. Anatomy of this phytomorph is fundamentally different from the phytomorphic tissue of *Archimedes*, *Lyropora*, and similar forms, as well as from the phytomorph of *Bicorbula* Condra and Elias from the Kaibab limestone—all of which we interpret as consortia of Bryozoa and phytomorphs, probably algae.

The lowest horizon known to us in which *Thamniscus* with encrusting phytomorph occurs is Brownville limestone, which is the highest rock of Pennsylvanian age in the Midcontinent. Below the Brownville only very few specimens of *Thamniscus*-like forms have been found, and none of these has the phytomorphic encrustation. We restrict the genus *Thamniscus* to the forms with phytomorphic encrustations. In this restricted sense the genus becomes apparently restricted to Permian and the very top of Pennsylvanian, and we interpret it as another case of bryozoan-algal consortium. The other species previously classified with *Thamniscus* and which are *Polypora*-like but have only few or no dissepiments and do not have encrusting phytomorph we exclude from *Thamniscus* (restricted).

RELIABILITY IN DETERMINATION OF BATHYMETRIC CONDITIONS OF DEPOSITION BASED ON FOSSIL FISH REMAINS

BY LORE ROSE DAVID

Studies of fish remains in core sections of numerous California oil wells, many of which show continuous coring, have revealed interesting data on the changes of fish faunas and the percentages of various forms in different parts of the sections.

Of special interest are the extensive bathypelagic faunas in the Upper Miocene Upper Mohnian section of all the wells in the Long Beach, Wilmington, and Torrance oil fields of the Los Angeles

Basin. A small number of forms is present: *Cyclothone* is extremely abundant representing 50 to 75% of all forms present, Myctophidae come next with 15 to 35%, usually represented by *Lampanyctus* sp. only; a large scaled form, probably *Myctophum*, is rarer; *Bathylagus* represents some 10% of the entire fauna, and the remaining forms as *Chauliodus* or undetermined Stomatoidea, Gadoidea as *Eclipes* and *Bathygadus* represent about 1% of the fauna. Findings of William Beebe (Zoologica, Vol. XXIX, 1944, p. 59ff.) on Recent bathypelagic faunas show very similar numbers. In the Pacific Ocean cyclothones form 54.7% of the entire catches, myctophids 36.9%, other fish 8.4%. The Myctophidae are represented by quite a number of species in various regions, but two of these in general represent 86 to 89% of the entire catch taken of the family. The similarity in composition of fossil and Recent bathypelagic faunas indicates the reliability of the fossil fauna in regard to bathymetric determinations.

True abyssal assemblages, to be distinguished from the above-mentioned bathypelagic faunas, should be mentioned in this connection. These are present in Pliocene and Upper Miocene Delmontian strata of the Los Angeles and Ventura basins. Representing Macrouridae in greatest numbers are the Bathygadidae, other Gadoidea, Myctophidae and Halosauridae rarely, and a few others, as characteristic for known Recent abyssal faunas.

EVOLUTION OF *MESOLOBUS* AND AGE OF THE EARLIEST *WALCHIA*

BY MAXIM K. ELIAS

At Bond and McCoy, central Colorado, species of *Mesolobus* occur at several horizons and provide abundant material for the study of variation and evolution of this genus.

The earliest form (basal Des Moines or basal Kanawha) has a striate (lirate) conch and an unstabilized, poorly developed median fold. In later forms, which are developed from it and are common in the Midcontinent, the conch remains lirate, while the median fold becomes emphasized and stabilized (*M. striatus*). In still later forms here and in the Midcontinent the conch becomes smooth, while the median fold becomes still more emphasized by a corresponding deep crenulation at the center of the shell margin.

In central Colorado another contemporaneous form, instead of becoming progressively smoother, developed coarser lirae than in the types of *M. striatus*. The new species with coarse lirae is of Marmaton (upper Des Moines) age and furnishes an apparent link between *M. striatus* and the Russian species described by Ivanov and Ivanova (1936) as *Chonetes mesolobus* N. and P., but which has even coarser lirae than the new species from Bond. The Russian species belongs to the Teguliferina beds or basal Gjelian of Moscow basin, which is correlated with the Missouri series of the Midcontinent.

The age of the *Walchia* flora which occurs at Bond below the new species of *Mesolobus* and above the zone of *M. striatus* is thus fixed as late Cherokee = early Carbondale = early Allegheny.

INTRASPECIFIC VARIATION IN THE ONTOGENY OF *PRIONOTROPIS WOOLLGARI* AND *PRIONOCYCLUS WYOMINGENSIS*

BY OTTO HAAS

Examination of about 450 specimens of *Prionotropis woollgari* Meek (? not Mantell) and about 40 of *Prionocyclus wyomingensis* Meek from the Western Interior of the United States proves a high degree of intraspecific variation and ontogenic change in these species. Interaction of both factors results in a bewildering diversity of morphologic aspects, particularly in *P. woollgari*. Both species are considered polytypic.

Meek's original (1876, pl. 7, figs. 1a, b) is designated lectotype of *P. woollgari* Meek; six varieties are separated from the typical form, and *P. branneri* Dickerson is considered as being a seventh. In addition to *Ammonites percarinatus* Hall and Meek, *A. graysonensis* Shumard is also included in the synonymy of Meek's species. The ontogeny of all those forms is closely followed with respect to dimensions, length of body chamber and full size attained, whorl section, ornamentation, and sutures. Intraspecific variation appears to be largely due to differentiation in speed of development.

White's (1880) specimens of *Prionocyclus wyomingensis* are recognized as being typical of this species; two varieties are separated. The ontogeny is studied along the same lines as that of *Prionotropis woollgari*, with similar results, especially in sutural development.

Both species being genotypes, the differences between them stand for those between their genera. They are found just sufficient to warrant, by present taxonomic standards, generic separation. One variety of *P. woollgari*, however, is somewhat transitional between both genera.

CALCAREOUS SPONGE IN DEVONIAN OF AUSTRALIA

BY B. F. HOWELL

Well-preserved calcareous sponges are very rare in pre-Carboniferous rocks. The discovery in the Upper Devonian Upper Sporadoceras beds of Western Australia of two new species that appear to be Calcispongiae is therefore worthy of note. These Australian sponges somewhat resemble in form the German Devonian species described by Spriestersbach in 1935 as *Scribroporella socialis*, but their body walls are much thicker than are those of the German form. The writer has been permitted to describe these sponges through the courtesy of Professor Curt Teichert, who collected them. They are believed to represent a new genus.

NEW GENUS OF TEREBELLID WORM FROM THE PENNSYLVANIAN OF TEXAS

BY B. F. HOWELL

Tubes of terebellid worms have been described from Jurassic and Cambrian rocks. Through the kindness of Dr. J. B. Knight the writer is now able to report the occurrence of a new genus and species in the Pennsylvanian of Texas. The new tube is remarkable for the perfection of its architecture; the walls are very carefully built of bits of shells and other calcareous materials.

ORDOVICIAN TRILOBITE GENUS, *CYBELE*, IN NORTH AMERICA

BY B. F. HOWELL

Some authors have doubted whether the European Ordovician encrinurid genus, *Cybele*, was represented in North America, contending that all the species from this continent which have been placed in that genus belong rather in *Cybeloides*. A re-examination of the types of "*Encrinurus*" *mirus* Billings and the discovery by Schuchert and Dunbar of additional specimens of that species in Newfoundland prove that *mira* is a true *Cybele*. This species, *Cybele spicata* Raymond, of the Collingwood formation of Ontario, and two examples of a new species of *Cybele* in the collection of Princeton University that were probably collected from the Trenton limestone of New York all indicate that the genus lived in North American Ordovician seas. Both *C. mira* and the new species have genal spines, so that character, which was thought by Slocum to be a distinguishing feature of his genus, *Cybeloides*, is found in some species of *Cybele* also, and the only characters which distinguish *Cybeloides* from *Cybele* are the two longitudinal furrows on its cranidium.

EUROPEAN TRILOBITE GENUS, *EUDOLATITES*, IN NORTH AMERICA

BY B. F. HOWELL AND J. T. SANFORD

Eudolatites, a genus of the Dalmanitidae proposed by Delo in 1935, has been known heretofore only from the Ordovician and Silurian of Europe. The discovery in the Silurian Sodus and Brewerton formations of New York of pygidia of two new species that appear to be referable to this genus is therefore of interest. These specimens, now in the paleontological collection of Princeton University, will be described by the authors.

CERATOCEPHALA IN GASPE AND NEWFOUNDLAND

BY CECIL H. KINDLE

Well-preserved specimens of the trilobite *Ceratocephala goniata* from the Silurian of Gaspe show this species to have three pairs of glabellar furrows and a fringe of spines around the lateral margin of the head. This makes it more similar to the European *C. verneuilli* than has been realized.

A new *Ceratocephala* has been found in the Table Head limestone of Newfoundland. This increases the known geologic range of the genus, for the oldest occurrence noted for it in *Index fossils of North America* is Upper Ordovician.

NOTE ON A POST-BUDA COMANCHE CRETACEOUS MICROFAUNA FROM THE EAST TEXAS EMBAYMENT

BY FRANK E. LOZO, JR.

Foraminifera and Ostracoda from an unnamed subsurface formation of bronze and gray shale between the Buda formation (Washita group, Comanche series of the Cretaceous) and beds of the Woodbine group of the Gulf Cretaceous have been studied. The fauna had definite Washita paleontologic affinities and is of Comanchean age. The fauna is characterized by a number of new species of Foraminifera in addition to many previously known from the Upper Washita.

NEW HAMILTON HEXACTINELLIDA

BY IRVING G. REIMANN

Pseudohydroceras is redescribed. Three new genera of small hexactinellids, rather common in the middle beds of the Ludlowville shale, are described. Small, soft, spherical carbonaceous bodies occurring in the sponges and in the matrix are suggested to be gemmulae.

NEW MIDDLE DEVONIAN OCTACTINELLIDS

BY IRVING G. REIMANN

Four new species of *Astraeospongia* Roemer are described, one from the Onondaga limestone and three from the Hamilton formation. *Astraeospongia hamiltonensis* Whiteaves (not Meek and Worthen) from the Hamilton of Ontario is renamed. A new genus from the Onondaga limestone, including five new species, is described.

NEW SILURIAN AND DEVONIAN BLASTOIDS

BY IRVING G. REIMANN

Three new genera distributed among the Codasteridae, Phaenoschismidae, and Pentremitidae are described. The genera *Troosticrinus*, *Pentremitidea*, and *Eleutherocrinus* are discussed, and certain morphological details redescribed or added. Earlier interpretations of the summit structures of *Troosticrinus* are questioned, and the genus is referred to the Pentremitidae. A divided posterior deltoid is shown to be characteristic of *Pentremitidea* and *Eleutherocrinus*. Two Silurian species of doubtful generic affinities, and 21 Devonian species are described.

REAL AND SIMULATED COLOR PATTERNS IN *MERISTELLA*

BY IRVING G. REIMANN

Radial color bands on *Meristella nasuta* (Conrad) from the Onondaga limestone, and two new species and a new variety of the same genus from the Hamilton formation are concluded to be the result of both pigmentation and light refraction or reflection caused by inner shell structures. On certain shells the pattern is strongly shown where the outer pigmented layer is absent.

PRELIMINARY ECOLOGICAL STUDY OF SOME RECENT PACIFIC OSTRACODA

BY W. THOMAS ROTHWELL, JR.

Considerable interest in the application of the Ostracoda to California Tertiary stratigraphic studies has been aroused by L. W. Le Roy's investigation of Pliocene and Pleistocene Ostracoda (Jour. Paleont., vol. 17, no. 4, 1943). My study is concerned with the ecologic relationships of recent Ostracoda as interpreted from 73 San Pedro Channel dredge samples collected by M. L. Natland and Alex Clark. The sample stations are located on a traverse from the littoral zone in the vicinity of Long Beach, California, to the beach at Avalon, Santa Catalina Island, California, a distance of 38 miles, with depths ranging from 1 to 2900 feet.

A graphic chart lists each station, the species, depth, abundance, degree of preservation, and in some cases the relative molt stage. Special emphasis is given to the companion occurrence of the Ostracoda with the Foraminifera since the depth distribution of the latter from the same samples

has previously been reported by M. L. Natland. (Univ. of Calif. Press, 1933). The results of the preliminary study suggest the following as important factors influencing ostracod distribution: (1) amount of clastic sediment in the water, (2) depth, and (3) biologic factors in common with calcareous algae, bryozoans, and foraminifers.

Genera recognized include *Bairdia, Brachycythere, Bythocypris, Cythereis, Cytherelloidea, Hemicythere, Loxoconcha, Paracypris,* and *Paracytheridea.* Two species previously reported fossil have been added to the list of those surviving to recent time—*Cythereis kewi* Le Roy and *Paracytheridea granti* Le Roy.

X-RAY STUDIES OF SURFACE LAYERS OF CRYSTALS: A REVIEW OF THE FIELD WITH SPECIAL REFERENCE TO QUARTZ

BY ELIZABETH J. ARMSTRONG

The reflection of X-rays from crystalline material depends very precisely on the orientation of the crystal structure with respect to the X-ray beam: the intensity of the reflection is related to the structural perfection of the crystal. Based on these fundamental principles various techniques have been and are being devised for the X-ray study of structural disturbances in artificially prepared crystal surfaces and natural faces. These techniques are reviewed, and their potentialities and limitations are noted, with special reference to their use in the study of ground and lapped quartz surfaces.

APPLICATION OF ELECTRICAL EARTH RESISTIVITY SURVEYS TO GROUND-WATER STUDIES IN OHIO

BY RALPH J. BERNHAGEN

Earth-resistivity surveys operate on the fundamental premise that earth materials of different character react differently to the passage of an electric current. Average apparent resistivities are measured to various depths at different stations suitably located to accommodate the purpose of the investigation. The interpretation of the data obtained in such a survey may be used to determine: (1) the thickness of glacial drift overlying bedrock, (2) the depth and extent of a body of water and nature of the underlying material, (3) the locations of ore bodies and placer deposits, (4) the location of possible water-bearing sand and gravel deposits.

Four electrodes are placed in a straight line in the ground at equal distances on a relatively plane surface; a commutated current is conducted into and through the ground between the two outer electrodes, and the voltage drop is measured between the two inner electrodes. The depth at which each reading is taken is determined empirically by the distance separating any two electrodes.

The Ohio Water Supply Board has inaugurated in the past year a program of resistivity surveys to assist in the location of ground-water supplies for Ohio's municipalities and industries, particularly in the glaciated area to delineate bodies of sand and gravel and to locate the channels of buried preglacial valleys. The interpretation of the resistivity data is considered in the light of all known geological information for the area under consideration; the favorable water-bearing conditions are outlined, and the locations for test drilling are recommended.

AN INTERPRETATION OF LOWER JAMES RIVER, VIRGINIA*

BY ROBERT O. BLOOMER

James River in Virginia, like other Appalachian rivers, has had a long complex geomorphic history. Certain peculiarities in the present course of the stream may be a key to its past history and may have some bearing on the interpretation of the regional geomorphology. Several changes in the average southeast course may be due to tilting, originating from an axis transverse to the southeast regional slope. These deviations from the southeast course may furthermore mark the location of former strands and thus indicate margins, at various times, of the Coastal Plain overlap.

Also the course of James River is such that there seems to have never been an advanced stage of the geomorphic cycle such as maturity or old age. A channel, however, in the Coastal Plain overlap below Richmond may have been formed by James River during Lower Cretaceous time. Thus James River may have been a large stream as early as the Cretaceous but never reduced to or near base level. Consequently the Tertiary peneplain postulated by some geologists may be doubted.

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AGING OF QUARTZ CRYSTAL OSCILLATOR PLATES

BY VIRGIL E. BOTTOM

Quartz plates of specific orientations are widely used to control the frequencies of radio oscillators. The physical properties of quartz make it a nearly ideal material for this purpose. Serious difficulty has been encountered with high-frequency plates due to aging. This has been shown to be due to deterioration of the surface of the plate. The effect is associated with the action of water vapor on a layer of misaligned quartz which results from the operations of grinding, lapping, and polishing. The remedy is removal of the material by etching. When quartz oscillator plates are etched by an amount sufficient to remove the misaligned material, the surfaces are apparently stable, and aging is negligible.

STAGES OF LAKE CHICAGO; THEIR CAUSES AND CORRELATIONS

BY J HARLEN BRETZ

The three stages of Glacial Lake Chicago record (1) the original lake level, (2) a time of rapid lowering because of outlet deepening, (3) a pause in this deepening and lowering, (4) renewed lowering because of renewed outlet deepening, and (5) a final level determined by a bedrock sill in the outlet channel floor. The dam for the lake was not the Valparaiso moraine, as earlier students have thought, but the younger Tinley moraine and its outwash. The valleyway across the Valparaiso moraine is much older than Lake Chicago. No stopping by retreat of rapids, either in glacial drift or in bedrock, was involved in the times of outlet deepening or the pauses between those times.

The outlet downcuttings are correlated with times when glacial lakes in the Erie and Huron basins discharged into Lake Chicago and greatly increased erosive action in its outlet channel. The two earlier static water levels record times when these eastern lakes discharged elsewhere and a boulder armor on the outlet floor brought deepening to an end.

INTERLACING OF PIEZOELECTRICITY WITH OTHER BRANCHES OF SCIENCE

BY W. G. CADY

To what extent is our knowledge of piezoelectricity and our ability to put it to work dependent on an acquaintance with the various sciences? Conversely, to what extent has piezoelectricity contributed to the physical or other sciences? An attempt is made to answer these questions. The topics considered will include a brief account of the nature of piezoelectricity, its relation to crystallography, and the methods for detecting or measuring the piezoelectric property; the use of thermodynamic methods for expressing the relations between piezoelectric, elastic, and dielectric phenomena; the interaction of these relations in the theory of the crystal resonator; the principles underlying the design of resonators, with regard to the various possible modes of vibration and their excitation and to the problem of making the vibrational frequency independent of temperature. Attention is directed to the use that has been made, in the progress toward this goal, of all branches of physics, but more particularly vibration theory, dielectric and electric circuit theory, polarized light, and X-rays; and in addition, geology, crystallography, and chemistry.

COMPOSITION OF THE GLACIAL DRIFT IN OHIO

BY G. W. CONREY

Two features of importance in evaluating the composition of glacial drift are the content of carbonates and the amount of clay. Dilute muriatic acid has been widely used in testing for the presence of carbonate. Effervescence indicates the presence of carbonates but not the amounts. Texture of the material can readily be estimated from a determination of the consistence (plasticity, stickiness, etc.). These characteristics are of great importance in studies of the genesis of Ohio soils.

Considering only the late Wisconsin drift, in western Ohio, where the underlying bedrock is chiefly limestone, the lime carbonate content varies from 25 to 45 per cent, the clay content from 15 to 24 per cent. In northeastern Ohio, where the bedrock is chiefly noncalcareous sandstone and shale, where the texture of the till is a loam, the drift may be noncalcareous to 6 or 8 feet below which there may be

2 to 5 per cent of lime carbonate border, where sandstones predominate. In contrast are the clay tills on the northern part of the belt where lime carbonate ranges from 1 to 18 per cent, and the clay content from 28 to 49 per cent. The lime carbonate is confined largely to the fine material. A consideration of the direction of movement of the ice suggests, as the source of the lime carbonate, (1) the basin now occupied by Lake Erie, (2) the limestone deposits in Canada north of Lake Erie, (3) lime carbonate precipitated from percolating waters.

Evidently it is necessary to recognize high-lime and low-lime drift. These weather to produce soils of markedly different characteristics.

CORRELATION OF SOME COALS IN SOUTHERN WEST VIRGINIA BY USE OF PLANT MICROFOSSILS

BY AUREAL T. CROSS

Extensive investigations are being made of the plant microfossils contained in the Lower Allegheny and Kanawha coals of West Virginia. The stratigraphic and geographic ranges are being determined in an effort to establish proper index values for these microfossils. The most readily identifiable plant residues, the megaspores, have been studied from about 35 coal horizons. Over 45 species, assigned to 14 genera, have been tentatively identified. These include more than 25 species of the large genus *Triletes* and several new species and varieties which await description. Investigation to date shows these spores to be fairly consistent geographically by their similar representation in the same horizons from each of several localities. These microfossils also furnish evidence of several major breaks in lithology of the strata involved. The most pronounced change in floristic composition occurs about the horizon of the Chilton "A" coal, which is between the Winifrede (Magoffin) and Salt Lick marine zones, or approximately in the position of the recently postulated boundary between the Allegheny and Kanawha series. A brief discussion is given concerning the methods used in the present investigation to obtain and prepare the samples for study.

STRUCTURE OF THE McCOY AREA, EAGLE AND ROUTT COUNTIES, COLORADO

BY HENRY F. DONNER

The area lies near the central part of the northwest quarter of Colorado and includes about 70 square miles in the transition zone between the Ancestral Rockies and the Central Colorado Basin. Overlaps and marked changes in lithology are very common.

Marked unconformities separate pre-Cambrian and Cambrian rocks, Mississippian and Pennsylvanian rocks, and Permian and Triassic rocks. Less conspicuous unconformities occur between the other formations.

Deformation took place probably in late Cretaceous time and produced open asymmetric folds. One set trending north-south with the steep limb on the west suggests forces from the east. Much smaller folds trending slightly north of northwest with the steep limb on the northeast side form the other set, are found only in the northern part, and are believed to be drag folds formed by the uplift in the northeast.

Faulting occurred shortly after folding, probably during late Cretaceous or early Tertiary time. The faults are all of the high-angle type. Most of them are normal faults and show considerable drag. In some cases reverse movement along the fault plane is indicated. Displacements of several hundred feet are very common.

Tertiary dikes and stocklike intrusives cut all older rocks. Miocene lavas have been tilted at angles up to 9°. Pliocene lavas have been uplifted but not tilted.

Erosion surfaces and terraces along the Colorado River testify to several uplifts during Quaternary time.

CLEVELAND MUSEUM COLLECTION OF FOSSIL FISHES FROM THE UPPER DEVONIAN OHIO SHALES

BY DAVID H. DUNKLE

A summary of vertebrate paleontology in the upper Devonian Ohio shales is presented. The development of the Cleveland Museum of Natural History Collection of fossil fishes from those

strata is specifically reviewed with discussions of the component fish types represented and their geologic and geographic distribution. Finally, a statement as to scope of the research program being conducted in this particular field by the Cleveland Museum is offered.

GEOLOGY OF CLEVELAND

BY DAVID H. DUNKLE

Physiographically Cleveland sits on the border between the Appalachian Plateau and the Central Lowland provinces, a boundary marked by the prominent but generally gently inclined Portage escarpment which traverses the immediate region from northeast to southwest. The Appalachian Plateau portion of the area is marked by high and rugged topography with narrow valleys and steep hills. Westward on the Lake Plain altitudes and relief are lower with long, rolling hills and glacial features such as eskers, moraines, and beach terraces. Drainage is northward into the Great Lakes system.

The region is underlain by Late Paleozoic rocks dipping southeast and exposing their uptilted edges in a relatively narrow belt paralleling Lake Erie. Certain formations are the source of a large and well-known fish fauna.

The following outline classification of the rocks to be observed within the city is offered for the convenience of the visiting members of Section E.

Pleistocene mantle
  Wisconsin and (?) Illinoian tills
    Unconformity

Pennsylvanian
  VI. Sharon conglomerate
    Unconformity

Mississippian
  V. Cuyahoga group
    Meadville shale
    Sharpsville sandstone
    Orangeville shale
  IV. Berea sandstone
    Unconformity

Devonian
  III. Bedford shale
  II. Cleveland shale
  I. Chagrin shale

The more important rivers have, in many places, cut into thick fillings of several large preglacial stream channels. These consist largely of laminated clays, especially in the lower parts where they are probably of lacustrine and fluviatile deposition. Their upper parts as well as the general surface of the whole area are covered by glacial tills of the Wisconsin and possibly remnants of the Illinoian stages.

CLASSIFICATION OF THE UPPER SILURIAN AND MIDDLE DEVONIAN STRATA OF THE MACKINAC REGION

BY GEORGE M. EHLERS

Two Upper Silurian and three Middle Devonian stratigraphic units, their lithologic and paleontologic characters, and their correlation with other rocks are noted.

The lower of the two Upper Silurian units, designated the Pointe aux Chenes formation, consists of green and red shale, gypsum, and some beds of dolomite and rests on the Engadine dolomite of Middle Silurian age. The upper part of the Pointe aux Chenes formation contains fossils characteristic of the Bertie formation of the Salina group. The upper formation of the Upper Silurian consists largely of thin, even-bedded, light-gray to cream-colored dolomites. Thick-bedded gray and buff dolomites, present in the upper part of this formation, have fossils related to species in the Akron dolomite.

The lowest of the three Middle Devonian units consists of dolomites and dolomitic sandstone and contains characteristic Oriskany fossils. The middle unit is composed of cherty dolomites and limestones with a lower Onondaga fauna. The highest unit of limestone, dolomite, and shale comprises the Detroit River group, the type exposures of which occur in southeastern Michigan and northwestern Ohio. South of the Straits region the Detroit River group is succeeded by the Middle Devonian Dundee limestone, Rogers City limestone, and Traverse group.

GEOLOGICAL INVESTIGATIONS OF UNITED STATES QUARTZ CRYSTAL DEPOSITS WITH SPECIAL REFERENCE TO THOSE OF ARKANSAS*

BY A. E. J. ENGEL AND OTHERS

Geological investigations of domestic deposits of quartz crystal were undertaken by the Geological Survey in the fall of 1942 in response to the urgent demand for crystals for war purposes. The work was continued throughout 1943, and during this period a rapid reconnaissance of all the more promising areas was made; literally hundreds of small deposits were examined, and the more noteworthy localities such as those of northwest Arkansas, the Piedmont districts of Virginia, North Carolina, and Georgia, and the placer deposits of Calaveras County, California, were studied in some detail. No deposits capable of producing large quantities of usable crystals were encountered. The most important reserves are those of western Arkansas, and this paper is concerned chiefly with a discussion of the geology and mineralogy of the Arkansas deposits

PHYSIOGRAPHIC DIVISIONS OF THE COLUMBIA PROVINCE

BY OTIS W. FREEMAN, J. D. FORRESTER AND R. L. LUPHER

The "Columbia Plateau," described by N. M. Fenneman in his *Physiography of the Western United States* (1931) should be renamed Columbia Province because of its diverse physiography—warped basins, high plains, plateaus, hills, and mountains—and its encirclement by high mountains.

The Columbia Province is characterized by: (1) lava flows (Tertiary to Recent) associated with continental deposits, (2) widespread warping, folding, and faulting, and youthful to mature erosion, (3) Columbia River drainage.

Four major divisions of the Columbia Province merit recognition as subprovinces rather than sections, from north to south: (1) Columbia Basin, (2) Central Mountains (Blue Mountains), (3) High Lava Plains, and (4) Owyhee Uplands.

The Columbia Basin is divided into six sections: (1) Deschutes-Columbia Plateau, (2) Central Plains (Pasco-Quincy Plains), (3) Yakima Folds, (4) Waterville Plateau, (5) Channeled Scablands, and (6) Palouse Hills. The lavas are of Tertiary age; the stage of erosion is largely mature.

In the Central Mountains Tertiary lavas and continental sediments lie discordantly on older igneous, metamorphic, and sedimentary rocks. The mountains represent a recently deformed and partly dissected lava plain. The suggested sections are: (1) Blue Mountains, (2) Wallowa-Seven Devils, and (3) Tri-State Uplands.

The High Lava Plains are dominantly composed of young, flat-lying basalts, recent cinder cones, and intervolcanic sediments. Drainage and stage of erosion are typically immature. The sections recognized are: (1) Upper Snake Plain, (2) Malheur-Boise Basin, and (3) Harney-High Desert.

The Owyhee Uplands are older lavas than those exposed in the High Lava Plains. Essentially the area is an unwarped plateau surmounted by the Owyhee Mountains and South Mountain old crystalline rocks. The Uplands are more deeply dissected than adjoining regions.

FREQUENCY ADJUSTMENT OF QUARTZ OSCILLATOR PLATES BY X-RAYS

BY CLIFFORD FRONDEL

A BT quartz oscillator plate irradiated with X-rays, or by certain other radiations, gradually becomes smoky in color, and the oscillator frequency decreases. There is little or no accompanying change in crystal activity. The total frequency change is limited and varies with the initial frequency (thickness) of the plate. Changes of the order of 500 c.p.s. to 3000 c.p.s. can be obtained in the

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frequency range from 5 to 9 megacycles. The rate of change of frequency is primarily determined by the intensity and wave length of the X-radiations. Both the rate of change and the total change of frequency increase with increasing initial frequency of the plate. Rates now achieved in production average about 40 c.p.s. per minute of exposure to X-rays.

The frequency change brought about by irradiation can be reversed and the plate restored to its original frequency and color by baking at temperatures over about 175° Centigrade. Irradiated plates are stable at lower temperatures. The plates also can be sensitized to irradiation by prior baking.

The irradiation technique presents a number of practical advantages in the manufacture of oscillator plates: (1) Extremely precise frequency adjustments can be made by oscillating the plate in the X-ray beam, following visually the frequency change on a meter until the desired value is reached. The adjustment can be effected, under suitable circumstances, while the crystal is oscillating in its permanent holder. (2) The frequency of stabilized crystals can be adjusted without disturbing the surface condition of the quartz. (3) The fact that the frequency change is downward permits the salvage of crystals that have been overshot in frequency during manufacture. Similarly, plates that have gone over frequency due to aging, recleaning, or underplating may be recovered.

TWINNING IN QUARTZ

BY CLIFFORD FRONDEL

Much of the raw quartz available to the oscillator-plate industry is useless because of Dauphiné (electrical) or Brazil (optical) twinning. Dauphiné twinning can be removed artificially from quartz; it also can be produced artificially by (1) cooling beta-quartz down through the 573°C. inversion point to alpha-quartz; (2) rapid cooling to room temperature from initial temperatures in the range from roughly 200° to 573°; (3) rapid heating from room temperature to temperatures in the range from roughly 200° to 573°; and (4) local application of pressure at room temperature or higher. Secondary twinning of this type is not accompanied by a permanent, geometrically defined distortion of the crystal and is distinct from twin gliding produced by mechanical deformation.

Additional types of secondary twins can be produced artificially by superposing Dauphiné twinning upon other kinds of twins already present. Of these, the so-called Combined law, produced by superposing Dauphiné upon Brazil twinning, is of special interest. This law was recognized by W. J. Lewis (1899) largely on theoretical grounds but has escaped mention in modern literature. Some natural examples of primary Combined twinning are mentioned.

The boundaries of natural Dauphiné twinning can also be moved artificially through a crystal. The strategy of detwinning Dauphiné twinned oscillator plates is based on heating the quartz over the inversion point to beta-quartz, in which Dauphiné twinning cannot exist by reasons of symmetry, and controlling the cooling conditions so that the re-inversion proceeds from one center rather than many. Dauphiné twinned oscillator plates homogenized by heating over 573°C. often show a "memory" phenomena—both the incidence and distribution of the secondary twinning which reappears on cooling are influenced by the original twinning. The secondary twinning usually exhibits a regular pattern dependent on the crystallographic orientation and outward shape of the oscillator plate and on the cooling conditions. The secondary twinning, related to cooling cracks, originates in strains.

Both cracking and secondary twinning in quartz are primarily controlled by the rate of heating and cooling. While both effects are especially favored at 573°C., neither is necessarily a concomitant of, or indicative of, inversion from beta-quartz.

IRON-ORE PORT OF CLEVELAND

BY M. D. HARBAUGH

The Port of Cleveland is one of the principal receiving points for lake-borne iron ore from the Lake Superior ranges, ranking first, in recent years, in tonnage handled over its several docks. More than 13.5 million gross tons, or 16 per cent of the total lake movement, was unloaded there in 1943, and a million tons more than that in 1942. Less than one-third of this is consumed in local blast furnaces; the rest is shipped overland to plants in the Ohio Valley.

The Pennsylvania Railroad dock, which is the largest, is located directly on Lake Erie. It handles ore solely for rail shipment. The other docks lie along the Cuyahoga ("crooked") River, from near its mouth to a point about 6 miles upstream. Except for the Erie Railroad dock, the river docks are alongside the several local blast furnace plants—*i.e.*, of the American Steel and Wire Co., Republic Steel Corporation, Otis plant of Jones and Laughlin, and a new D. P. C. furnace operated by Republic.

The distinctive, long ore carriers tie up at these docks, and batteries of clam-shell dipper unloading machines, all working simultaneously over different sections of the vessel hold, scoop up 5 to 17 tons of ore at a time and unload a cargo of 8000 to 15,000 tons or more in a few hours.

QUARTZ CRYSTAL UNIT

BY R. A. HEISING

The need for crystal control in radio communication is discussed. The manner in which the crystal fits into the radio circuits is shown. Several specific crystal units are described, and the principal physical and engineering factors that influenced the designs are outlined.

ERRORS IN THE SYNOPTIC WEATHER CHARTS WHICH COVER THE GREENLAND REGION

BY WM. H. HOBBS

Though it has been proven that cyclones which approach Greenland from the west are halted at the west coast and pass up Davis Strait, the Daily Weather Charts of the British Air Ministry continue to represent them as crossing Greenland exactly as though it were an expanse of sea. The data which have been regularly issued from Greenland coastal stations by radio have been quite inadequate and even unreliable. The true conditions have been revealed by special Greenland expeditions and confirmed by "post-mortem" studies for the special years 1930–1931 and 1932–1933.

RHYTHM IN THE GREENLAND GLACIAL ANTICYCLONE

BY WM. H. HOBBS

The cold interior region of the Greenland glacial anticyclone, within which air masses are descending, was occupied by both German and British meteorological stations during 1930–1931, at the former for a full year and at the latter for 7 winter months. An approach to periodicity or rhythm is betrayed in the records of all the meteorological elements, which surprisingly vary together.

MAJOR BURIED VALLEY IN EAST CENTRAL ILLINOIS AND ITS REGIONAL RELATIONSHIPS*

BY LELAND HORBERG

A large buried valley, here named Mahomet Valley, extends westward from the Illinois-Indiana line through central Illinois and enters the well-known bedrock valley along Illinois River formerly occupied by the ancient Mississippi. Mahomet Valley is considered preglacial, because Kansan, Aftonian, and possibly Nebraskan deposits occur within the channel. By the end of Illinoian time the valley was so completely filled with drift that the Sangamon interglacial plain continued across it without interruption. Continuation of the valley eastward into Indiana is suggested by well records and outcrop data. Diversion of drainage from Mahomet Valley into the lower Wabash bedrock channel was probably caused by the advance of Kansan ice.

It is proposed as a new working hypothesis that Mahomet Valley represents the lower course of Teays River, a master preglacial stream which probably headed near the eastern scarp of the Blue Ridge in North Carolina and flowed westward across Ohio, northern Indiana, and central Illinois, finally discharging into the Gulf Embayment through bedrock valleys now generally occupied by the present Illinois and Mississippi rivers.

STRATEGIC SIGNIFICANCE OF ALASKA

BY GEORGE D. HUBBARD

Alaska with her Aleutian Islands and the Alexander Archipelago sits across the north Pacific as a crown. Its position almost cutting off, with the tip of Siberia, communication between Pacific

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and Arctic oceans has been a guard to check polar fronts and cold air from rolling farther south and a more effective barrier to the Japan Stream, thereby holding its warm waters to bathe the American coast.

Its history suggests the importance attached to it by English and Dutch in the eighteenth century and by Russians in the eighteenth and nineteenth centuries.

Alaska stands between Russia and Japan on the west and the United States and Canada on the east. Politically and militarily it has absorbed many shocks as a buffer area.

Economically it stands on a Pacific border where its forests, furs, fish, and, some day, its coal resources may be put to good use.

Commercially it is on a great circle highway for trade and travel between Orient and Occident. Vessels plying between Yokohama and Vladivostok over there and Vancouver and Seattle on our side have to turn out for the Aleutian Islands, so its products are near a great east-west highway. Perhaps equally significant is its position as a station on air routes over the pole between western United States and northern Russia; or between Japan and Siberia, and industrial centers on the Atlantic in the United States and Europe.

Developments in Alaska for aviation point to its importance today as a strategic military position and strongly suggest future commercial possibilities.

ELKHORN ABANDONED CHANNEL OF THE KENTUCKY RIVER

BY WILLARD ROUSE JILLSON

Fluviatile gravels were recognized in northeastern Kentucky by Sidney S. Lyon in 1857. In 1877 Nathaniel Southgate Shaler commented the high level bordering the deep rivers of Kentucky. In 1888 Marius R. Campbell named them the Irvine formation, and in 1914 Arthur M. Miller fixed their age as Pliocene.

In 1941 the present writer discovered these patchy high-level gravels filling an old rock cut of the Kentucky River at an elevation of about 750 feet along South Benson Creek some 4 miles west of Frankfort. He subsequently mapped and described as the "Benson Channel" the first extensive buried Pliocene channel of the Kentucky River about 20 miles in length and some 250 to 300 feet above the present river. The same year he discovered a continuation of this fossil channel—also some 20 miles in length—some 25 miles to the north and in 1944 described it as the "Pot Ripple Channel."

The course of the Pliocene Kentucky River between the "Benson" and the "Pot Ripple" channels is the central theme of this paper. The field work as on each of the mapped areas was done by the writer and completed during June 1944. This new segment, principally along the waters of Main Elkhorn Creek, has been designated the "Elkhorn Channel." Maximum entrenchment below the level of the mid-Tertiary peneplane is 150 feet. Generally throughout its course of about 16 miles the depth of its rock-cut channel, the thickness of the channel "fill" of Irvine gravels, and the levels of the old flood or gradation plane—750 to 785 feet A.T.—correspond and unite with the adjoining upper "Benson" and the lower "Pot Ripple" channels to give a total length of high-level, Pliocene abandoned channel of the Kentucky River—from Tyrone to Lockport—of about 55 miles. Throughout this distance the "channel fill" of Irvine sand, clay, and gravel were but slightly touched, as in interfluve areas by post-Pliocene erosion which ranged from 50 to 75 feet.

ORIGIN OF MACKINAC BRECCIA

BY KENNETH K. LANDES

The Mackinac breccia occurs in vertical masses, which transect all formations between the lower part of the Upper Silurian Pointe aux Chenes formation and the Middle Dundee limestone. The rock between the breccia masses is tilted and faulted and is termed megabreccia.

The theory is developed that the collapse of salt caverns around the rim of the Salina basin in pre-Dundee time produced both the breccia and the megabreccia.

SLOANS VALLEY CAVERN SYSTEM, PULASKI COUNTY, KENTUCKY

BY CLYDE A. MALOTT

The Sloans Valley cavern system is an integrated system of well-developed subterranean conduits dominated by the discharge of the in-fed surface waters from a well-developed karst valley within the western edge of the Cumberland Plateau along the Cumberland River, 5 miles southeast of Burnside, Pulaski County, Kentucky. The main linear route is fully 3½ miles in length, which, through a descent of 160 feet, traverses six Mississippian limestone formations from the Glen Dean above to the St. Louis below. The cavern system is mainly developed along two sets of joints and descends in an oblique up-dip north-northwest direction for approximately 2 miles and then obliquely down-dip east-northeast to the Cumberland River, deviating surprisingly from the surface valley. The carefully mapped underground course connects with six cavern entrances and has unmapped subterranean connections with at least five others in the perched valley floor. The system has been progressively developed in five successive sections. The older sections are composed of four or more much collapsed tiers through a range of 125 feet, while the latest developed section is a simple joint-determined course. The system is wholly the work of the local drainage from less than 10 square miles, as no waters enter it from outside the drainage basin. The in-flowing surface waters have made this great cavern system in lieu of deepening the vacated surface valley for as much as 160 feet below its perched karsted floor.

QUARTZ AS A MINERAL RESOURCE

BY ROBERT B. McCORMICK

The recent growth of the quartz-crystal industry is traced, and comparisons are made to show the relative size of the quartz crystal and other nonmetallic mineral industries. The equipment used for the inspection and grading of quartz crystal at the National Bureau of Standards is described. A detailed discussion of inclusions and other imperfections found in quartz crystal is presented.

FLUCTUATIONS IN THE LEVEL OF LAKE MICHIGAN AND THEIR RELATION TO SHORE EROSION

BY WILLIAM E. POWERS AND EDWARD C. DAPPLES

Unusually high levels of Lake Michigan during the summer of 1943 accompanied damaging wave erosion at many places along its west shore. On the Northwestern University Campus in Evanston, wave action severely damaged most of the groins. Much of the beach deposits were removed and the adjacent cliff was eroded for the first time since 1930. During the dry 1930's, lake levels were generally low and beaches broad; wave erosion occurred. In the high stages of 1929–1930 and 1918 the shore was severely damaged.

An analysis of the records for Lake Michigan suggests cyclical recurrence of high and low levels. Periods of high water recur at intervals of approximately 11 years, though there is some support for a cycle of about 23 years marked by a maximum high stage, a minimum low, two intermediate lows, and two intermediate highs. These fluctuations are related to variations in precipitation, evaporation loss, storage in the lake, and other factors. The hydrologic equation is difficult to set up for the Great Lakes, because evaporation is imperfectly understood, and the underground seepage cannot be measured directly.

Although the rising levels in 1942 and 1943 indicated an even higher mean level in 1944, the records of the first half of 1944 are lower than expected. It is clear, however, that periods of high mean lake level and excessive shore erosion will recur at periods of 11 years or less. Until better data are made available beach engineers should prepare protective structures to meet the awaited wave attack.

SOME EVIDENCE REGARDING GEODE FORMATION

BY PERCIVAL ROBERTSON

Study of the sphericity and degree of hollowness, as well as some surface characteristics and mineral composition of geodes found in the Warsaw beds of Illinois and Missouri, has led to the conclusion that a theory involving the formation of the geode essentially simultaneously with the

deposition of sediments which formed the beds is desirable. Laboratory experimentation has shown that shells of colloidal silica can be made starting with sodium silicate syrup by processes not greatly unlike natural ones, except that sodium silicate syrup is at present unknown in seas or their tributaries. The occurrence of quartz crystals terminated by single rhombohedrons is not uncommon in these geodes, and the relationship of these crystals to layers of botyroidal chalcedony is pointed out.

UNDERGROUND STORAGE OF NATURAL GAS IN WAYNE, SUMMIT, AND STARK COUNTIES, OHIO

BY J. J. SCHMIDT AND J. E. SCHAEFER

Underground storage of natural gas has been practiced by the natural gas industry for many years, but it has been greatly extended in Ohio during the past 6 years.

The unprecedented requirements for natural gas occasioned by the war were partially responsible for the inauguration of many storage projects. Nevertheless, it is doubtful if storage on such a large scale would have been resorted to had it not been for the "know how" which was provided in the 10-year period prior to the war as a result of the technical evolution which has gradually taken place in the natural gas industry. Production engineering methods activated by geological thinking provided the tool whereby storage projects could be planned out and entered into safely without the fear of losing the large and valuable quantities of gas which must be returned to the earth as the first step in underground storage.

The selection of a proper pool to be used for gas storage requires considerable study. Some of the most important factors are (1) size of pool, (2) sand conditions, (3) original and present rock pressures, (4) original open flows and production of wells, and (5) type of storage desired.

The gas-storage operations of The East Ohio Gas Company in Wayne, Summit, and Stark counties are discussed.

DIOXAN IN THE MINERALOGICAL LABORATORY

BY M. W. SENSTIUS

This clear "water-white" alicyclic ether—diethylene oxide—has been used in the author's laboratory for the last 3 years with satisfactory results (1) as a diluent for heavy liquids (constant sg = 1.03); (2) as a clean index liquid (n = 1.42); (3) for displaying oily immersion liquids; (4) as a solvent for Canada balsam and synthetic resins.

These uses arise from its remarkable property of mixing freely with water and with a great variety of organic substances, oils, waxes, and resins. Thus tetrabrom ethane can easily be recovered from mixtures with dioxan by fractional washing with water, the last traces of water being removed with inert desiccants, such as commercial silica gel, "Drierite" ($CaSO_4$), calcium chloride, ignited Epsom salt ($MgSO_4$), or others. *It should not be recovered by fractional distillation for fear of explosion.*

Unlike benzene, nitrobenzene, and carbon tetrachloride, the faintly aromatic vapor of dioxan is nonpoisonous, while its rate of evaporation (b.p. = 101.1 C) is far less than that of acetone, alcohol, benzene, or carbon tetrachloride.

COMPARISON OF SEDIMENTS FROM LAKE ERIE AND LAKE MICHIGAN

BY M. W. SENSTIUS AND JOHN DE SISTO

Quantitative data are presented on the mechanical, chemical, and mineral composition of sediments from around the Bass Islands group of western Lake Erie and from various localities in Lake Michigan. The samples were obtained from known depths with a Petersen dredge.

While the mechanical and chemical compositions differ, the mineral contents are essentially similar. Because of the characteristic occurrence of celestite in the parent rocks of western Lake Erie, a special search was made for this mineral as a possible guide to the course of migration of the sediments. Celestite was not found in the sediments. Apparently the solubility of strontium sulfate is still too great for its preservation.

UPPER DEVONIAN NEAR-SHORE SEDIMENTATION IN WESTERN PENNSYLVANIA*

BY R. E. SHERRILL

The Catskill-Chemung facies relationship is well known through the work of Barrell, Willard, Caster, and others. During the past 8 years the Pennsylvania Geological Survey has been mapping in detail the oil-producing sand bodies which occur in the transition zone from continental Catskill to marine Chemung. Ancient beaches, off-shore bars, and channel deposits are recognized at several horizons. Periods of rapid subsidence, during which the shore retreated southeastward and permitted preservation of the sand bodies, alternated with restive stages during which continental deposition drove the shore northwestward. In general continental deposition predominated, and the average position of the shore in the upper sands, like the Venango, is northwest of its position in lower sands, like the Warren.

GEOLOGY AS AN AVOCATION

BY G. WINSTON SINCLAIR

Education of discharged men and women as well as ordinary students must consider the needs of those who will never be professional geologists, but who may find in the science a satisfying avocation. Such citizens can make real contributions to geology, and geology can be very rewarding to them. There are many ways in which geologists as teachers and individuals can encourage amateur work in the science and direct it into paths of greatest usefulness. They should be aware of the need and opportunity.

OSTRACODA OF THE OLENTANGY SHALE, FRANKLIN AND DELAWARE COUNTIES, CENTRAL OHIO

BY GRACE ANNE STEWART AND WILLIAM EDWARD HENDRIX

Although ostracodes have been recognized in the Olentangy shale from time to time no systematic search was made for them until 1939. At that time the writers collected 29 samples of shale from various outcrop areas along the Olentangy River and its tributaries in Franklin and Delaware counties, throughout a distance of about 25 miles. The shale was broken down, washed and sieved, and examined. Thirteen samples contained ostracodes, some only one or two species, but others a fairly rich fauna.

The fauna consists almost entirely of undescribed forms. The 29 species identified are distributed through 11 families and 15 genera, with 5 genera recognized for the first time. With the exception of two species, *Ctenoculina cicatricosa* (Warthin) and *Bythocypris lucasensis* Stewart, the entire species assemblage is new.

The fauna is definitely not middle Devonian, since characteristic forms such as *Ponderodictya unicornis* Van Pelt and others are not present. The genus *Richterina* suggests an upper Devonian age, and Mississippian affiliations are found in the several species of *Bairdia* and *Amphissites*, genera typically Mississippian but not entirely restricted to that period. Several representatives of Glyptopleuridae, a family not previously represented below the Mississippian, were found.

At present we cannot correlate this fauna definitely with any known fauna in the New York section, but it should aid in determining the age and stratigraphic relations of the Olentangy shale.

OSTRACODA OF THE PLUM BROOK SHALE, ERIE COUNTY, OHIO

BY GRACE ANNE STEWART AND WILLIAM EDWARD HENDRIX

Microscopic fossils in the Plum Brook shale have not received much attention apart from the conodonts which were described by Stauffer in 1938. Ostracodes have been recognized at various times, but no particular attention was paid to them. The writers began a study of them in 1939 so that a comparison with those in the Olentangy shale in central Ohio could be made.

Twenty-two samples of shale were collected, carefully broken down, and examined. Ten samples contain ostracodes from which 19 species were identified. Although the fauna is not prolific we have a more complete picture of these organisms than was known before.

* Given with the approval of the Pennsylvania State Geologist.

The Hamilton age of the ostracodes is determined readily by such species as *Ponderodictya unicornis* Van Pelt, *Ulrichia conradi* Jones, and *Poloniella cingulata* Warthin. Likewise, the resemblance to the Silica shale ostracode fauna is striking. Five species have been tentatively identified with forms in the Olentangy shale at Lewis Center Run in Delaware County. The latter are poorly preserved, and their proper identification is open to question.

Three new genera (*Punctoprimitia*, *Bythocyproidea*, and *Ortonella*), seven new species, and one new variety have been recognized.

ORES FOR THE FUTURE AND USE OF THE SPECTROGRAPH IN SEEKING THEM

BY LESTER W. STROCK

Much has been written recently concerning the depletion of high-grade metallic ore supplies. Present reserves of high-grade ores are dwindling to the danger point, and America, in particular, is approaching a point of nonsufficiency in regard to minerals previously mined in ample quantity. Mining men and teachers alike recognize the necessity of increased prospecting for mineral deposits; all emphasize that new techniques and attitudes will have to accompany renewed prospecting efforts.

The opinion is expressed that the mining and metal processing industries should support immediate activity, as a joint and co-operative effort, in searching for and developing the inevitable extension of low-grade ore deposits destined to supply our metals for the future. The value of the spectrograph as a prospecting tool in connection with such a program is outlined. Concrete evidence of its special advantages in discovering and exploring new types of ore deposits is briefly illustrated by outlining the part it played in the recent discovery of the large low-grade beryllium deposit at Iron Mountain, New Mexico.

SOME FEATURES OF THE SHARON CONGLOMERATE AND SANDSTONE IN NORTHEASTERN OHIO

BY MYRON T. STURGEON

The Sharon conglomerate and sandstone, the oldest member of the Pennsylvanian system in northeastern Ohio, rests disconformably on Mississippian formations. It was deposited by south-flowing streams in wide and deep valleys on a subsiding land surface. Its present discontinuous distribution results not only from original deposition along valleys but also from the erosion of much of the original deposits.

The Sharon is normally massive interbedded sandstone and conglomerate and has a maximum thickness of 200 feet or more. It is commonly the cap rock of the higher hills or ledge rock along the valleys. The ledges though massive show much stratification and cross-bedding, differential weathering, joints, caves, and an occasional cut-out and filled channel. The cross-bedding is almost without exception toward the southeast, south, and southwest. Ripple marks are present at very widely scattered localities.

The mineral composition is almost entirely quartz. The pebbles and cobbles range up to almost 6 inches, and all are rounded and commonly flattened or elongate. The sand grains range from fine to coarse, and usually fine to medium, and are subangular. Varieties of pebbles include transparent, white, yellow, pink, red, gray, black quartz, and quartzite, but the vast majority are milky quartz. The cement is silica or iron oxides. Other minor mineral constituents are mica, clay, and marcasite. The marcasite is present only locally in the basal 2 to 6 inches. The few fossils are plants, and the more common of these are *Calamites*.

QUARTZ FOR RADIO USE—1944

BY ALLYN C. SWINNERTON

The decision by the Signal Corps to utilize quartz-crystal units in many tactical types of radio sets for frequency-control purposes led immediately to many problems of raw-material supply and of production. In the course of the continuing solution of these problems much general interest in the piezoelectric use of quartz has been aroused, and some new knowledge regarding quartz has come to light.

To provide a background for the technical papers on quartz, this introductory paper outlines the commerce in quartz from the mines to the services, indicating the several government and private

agencies involved and their interrelationships. The current technology of quartz-plate manufacture is discussed briefly to indicate the relationship of the physical and mineralogical properties of quartz to the steps in manufacture of a unit, which meets the requirements of the armed service.

GLACIAL SLUICEWAYS AND LACUSTRINE PLAINS OF SOUTHERN INDIANA

BY WILLIAM D. THORNBURY

Numerous lacustrine plains of both Illinoian and Wisconsin age occur in southern Indiana, south of the Shelbyville moraine. The Illinoian plains mark the former sites of marginal lakes formed by the damming of the westward-flowing drainage lines against the Illinoian ice front. They are restricted to the western half of southern Indiana.

A more extensive system of lacustrine plains of Wisconsin age occurs along the valleys tributary to the streams which acted as sluiceways for Wisconsin melt waters. These flats are particularly well developed in the tributary valleys of the Wabash, Ohio, and the East and West forks of White River.

The Wisconsin lakes resulted from the damming of the tributary valleys by valley trains built down the courses of the streams which acted as Wisconsin sluiceways, resulting in the aggradation of the tributary valleys to the levels of the valley trains. The altitudes of the lacustrine plains descend regularly downstream with the slopes of the valley trains.

While the fill in the main valleys is largely sand and gravel, in the tributaries it is calcareous silt and clay. Calcareous concretions are usually common near the junction of the tributary with the main stream but usually are scarce or absent toward the heads of the lacustrine plains.

Along the lower Wabash and Ohio valleys extensive deposits of dune sand and loess of Wisconsin age were derived from the valley trains and lake flats at times of low water.

QUARTZ CRYSTAL DEPOSITS OUTSIDE THE UNITED STATES

BY CARL TOLMAN, ROBERT D. BUTLER AND RICHARD E. STOIBER

Almost all the quartz crystal used for piezoelectric purposes by the United States and Great Britain is obtained from Brazil. Russia produces some quartz. Very small amounts come from Australia, Colombia, Guatemala, Madagascar, and the United States. Even samples of usable material from other countries are rare. Most of the available geological information on quartz deposits in foreign countries is contained in engineering reports by field engineers of the Foreign Economic Administration; few specific detailed geological studies have been made.

Production of crystals in Brazil, largely in Minas Geraes, Goiaz, and Bahia over an area comparable in size to half the United States, is remarkable in the light of the restricted occurrence or absence of such crystals insofar as known in other parts of the world. Innumerable deposits are productive in Brazil, and from them the average production is very small. The production of the largest consistent producer, which is very much in a class by itself, is of the order of 1 to 2 tons per month of eye-clean quartz. In any deposit the usable crystals represent an extremely small proportion of the quartz present, and estimates of the ratio of the volume of material moved to eye-clean crystal produced varies from 3500 to 1 to as high as 100,000 to 1. Production from most deposits is from shallow depths by hand operations. Recently an increasing proportion of quartz is being obtained by using improved mining methods. Most production from other foreign countries is also from shallow depths.

The production of crystals is dominantly from quartz veins or from residual and alluvial deposits, presumably derived from the weathering of such bodies. The distribution of quartz deposits bears no observable relationship to other mineral deposits. The usual country rock is flat-lying shale or sandstone, but some veins in Brazil and elsewhere are found in schists or other metamorphic rocks and in igneous rocks. Generally, structural control is indicated by faults or shear zones or contacts between igneous and sedimentary or metamorphic rocks. Characteristically, the only recognizable vein material associated with piezoelectric quartz is cloudy or milky quartz except for the oxides of iron or manganese. However, in many places the matrix or country rock has been greatly weathered so that the original mineral content cannot be determined.

The characteristics of most of the vein deposits of quartz crystals of piezoelectric quality are those

of hypogene deposits formed under conditions of moderate temperature. In certain deposits the derivation of silica from surrounding silicious sediments has been suggested.

INFLUENCE OF GEOLOGY ON THE ECONOMY OF CLINTON COUNTY, OHIO

BY JAMES L. VANCE

Clinton County is located in the drainage basin of the Little Miami River in southwestern Ohio. The bedrock is Late Ordovician and Early Silurian dolomite, limestone, and shale. The county is on the east limb of the Cincinnati Arch, from which the strata dip only a few feet per mile. Exposures of bedrock are few due to glacial drift mantle 5 to 200 feet thick. The Cuba moraine and the Bloomington moraine, terminal moraines of Early and Late Wisconsin, pass through the county. South of the Cuba moraine the Illinoian drift has been covered by an ash-gray loess 3 to 10 feet thick. The section underlain by the Niagara and Brassfield formations are better drained than that underlain by the Richmond.

The Niagara dolomite is the only mineral deposit worked in the county. The county has a corn-hog economy; wheat is second in importance. Two regions, on the basis of soil fertility, soil types, land value, and general farming methods, are separated approximately by the Cuba moraine. Soils formed south of the Cuba moraine on the loess are plansols containing an impervious layer and requiring the use of surface drainage by means of shallow furrows.

The economy of Clinton County is not unlike that found in the surrounding counties in the Corn Belt. The two economic regions in the county are definitely a result of geologic conditions.

CONTINUING DEVELOPMENT OF QUARTZ-CRYSTAL UNITS

BY KARL S. VAN DYKE

The outstanding progress in the field of quartz crystals for communication use during the past few years has been in the development of machines and techniques for the quantity production and testing of crystals for which the basic research had largely been completed some years before. The quality of crystal units produced has increased directly as the circuits, and devices used for testing crystal performance have been able to indicate defects. The development of the vast new producing facility, in particular its ability to find new ways to fabricate crystals which would meet progressively higher standards as the tools to test for these have become available, is an amazing record of the ingenuity of American industry.

A brief account of the progressive improvement of crystal units is given, with particular attention to the development of methods for testing crystal performance. The basic radio circuit requirements for crystal performance are reviewed and possible directions of future development indicated.

CLIMATIC MAPS OF GEOLOGIC INTEREST

BY STEPHEN S. VISHER

As climatic data accumulates with the passing of the decades, maps of the annual average precipitation and temperature can now be supplemented by maps of several other climatic conditions. This paper presents 25 climatic maps of the United States, based on all the data available until recently. The regional contrasts in the climatic conditions mapped correlate with contrasts in various geologic agencies including weathering, erosion, transportation, and deposition. Some of the correlations are pointed out.

PERCENTAGE FREQUENCIES OF VERY DRY AND VERY WET SEASONS

BY STEPHEN S. VISHER

Supplementary to the study with 24 maps of the percentage frequencies of very dry and very wet months, published in the *Monthly Weather Review* for March 1944, a study has been made of the frequencies of wet and dry seasons. Maps of the four seasons were shown, as were maps of the maximum and minimum average precipitations for each season. (These maps are based on State averages, during 55 years.) The usual and extreme range or variation in seasonal precipitation are also mapped.

SOME POST-GLACIAL TIME DETERMINATIONS FROM LAMINATIONS IN SEDIMENT

BY IRA T. WILSON

The region of Sandusky Bay on Lake Erie was covered with glacial waters from the beginning of Lake Maumee to the end of Lake Elkton. The face of the ice sheet retreated about 250 miles during this interval. A core of sediment from near the edge of Sandusky Bay showed 12,223 varves. This gives an average rate of retreat of 108 feet per year for the ice assuming one varve was deposited per year. The cores were torn apart lengthwise, and the different varves parting at slightly different places made a serrated edge that facilitated counting. The summer layer is composed of fine quartz crystals and the winter layer of clay and silt. There is a decrease in the average number of varves per inch from the bottom of the deposit toward the top. The minimum number of varves per inch is 12, the maximum 38.

Douglas Lake in northern Michigan became an isolated warm-water lake with the lowering of Lake Algonquin to the Nipissing level. Most of the sediment subsequently deposited is an organic jelly. South Fish-Tail Bay is so situated that most of the diatoms growing in the lake have settled in layers interspersed with layers of organic matter. In the temperate zones two pulses of diatoms occur a year, one in the spring and one in the fall. In a 32-foot core of this sediment there were 102,692 layers of diatoms. On the supposition that two layers have been laid down per year the post-Algonquin period has been 51,346 years. The diatom layers were revealed by microscopic examinations of sections of the sediment cut on a microtome after the material had been infiltrated and embedded in paraffin.

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